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SPILITIC ROCKS OF SOUTHEASTERN NEW BRUNSWICK[1]

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ABSTRACT

Spilite flows and related keratophyric flows with associated pyroclastics of spilitic composition occur in the Caledonia ridge or massif of southeastern New Brunswick. The spilite flows are femic rocks or greenstones containing a sodic plagioclase which is generally albite. The associated tuffs are of similar composition, and the related keratophyric flows are soda-rich salic rocks. It is believed that the primary feldspar of the spilite was an acidic plagioclase containing about 15–30 per cent anorthite. The albite now present in the spilite is of secondary genesis, and there is some evidence that it was formed during the final stages of consolidation of the flow and during the period of extrusion of the volcanic series; that this alteration was due to soda-rich residual solutions in the consolidating flow, rather than to solutions from an outside source; and that the spilite magma was inherently rich in soda and different from the normal basaltic. A comparison with other occurrences of spilite shows a similarity in petrography and in associated rock types. The literature dealing with the spilite problem is briefly reviewed.

INTRODUCTION

Though spilitic rocks are of rather common occurrence in Europe, particularly in the British Isles, no North American occurrence of spilite apparently has been described. It is hoped that the following notes on the New Brunswick rocks may serve to incite discussion of this interesting rock type in North American literature.

GENERAL GEOLOGY

The northeastern end of the Caledonia Ridge or massif in Albert County, New Brunswick, the area covered by a horizontal ruling in

[1] Published with the permission of the Director, Geological Survey, Department of Mines, Ottawa, Canada.

the accompanying sketch, was mapped by the writer in 1930 and 1931. The rocks forming the massif are of pre-Carboniferous (pre-Cambrian?) age and comprise a bedded succession of spilite, keratophyre and quartz keratophyre flows, spilitic tuffs and tuffaceous sediments, phyllite and quartzite conglomerate, intruded by plutonic bodies of quartz diorite and albite granite, and by later dikes of augite andesite. The bedded succession appears to be structurally conformable throughout and has a probable thickness of at least

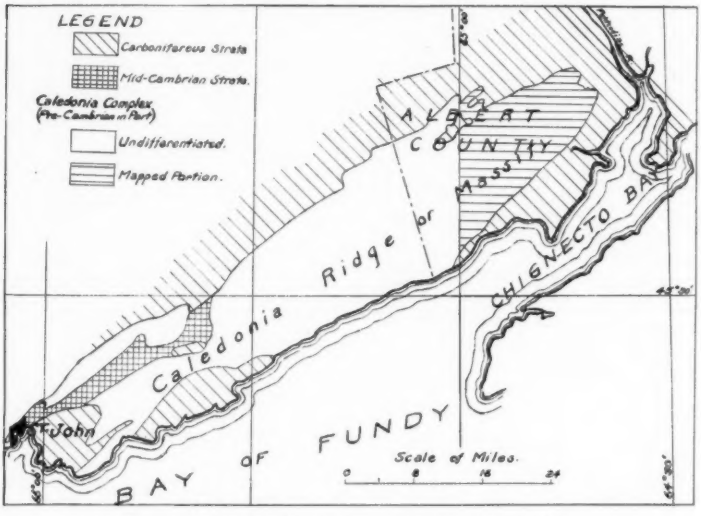

FIG. 1.—Spilitic rocks of southeastern New Brunswick

18,000 feet. The rocks are believed to be folded into a broad, open anticline with axis trending northeast for the most part, parallel to the ridge, and having moderate dips of 30° to 60°, the steeper dips prevailing farther from the anticlinal axis.

Dynamic metamorphism has affected all of the rocks of the massif, including the later plutonic intrusives. The spilite flows are commonly massive or locally slightly schistose. The keratophyre flows and the conglomerates are more strongly sheared, and the tuffs and tuffaceous sediments are generally schistose. The quartz diorite in part and the granite to large extent have been sheared to produce a

foliation structure in these rocks. A low grade of anamorphism has been imposed by dynamic metamorphism in the more thoroughly sheared rocks, producing phyllite and sericite schist from argillaceous sediments, chloritic schists from spilitic tuffs, and chloritic-sericitic schists from tuffaceous sediments.

Near Saint John, volcanic rocks lithologically similar to those of the area studied are overlain unconformably by mid-Cambrian strata, and it is probable that the greater part of the massif is composed of pre-Cambrian rocks.

SPILITE

Spilite flows occur intercalated with pyroclastics, normal sediments and keratophyre flows, throughout the bedded succession of these rocks, and form also the predominant part of two groups, one composed of spilite together with flows of altered basalt and beds of schistose basic tuff, the other composed almost wholly of spilite and quartz keratophyre flows interbedded as single flows a few feet in thickness and in bands (probably including several flows) more than 500 feet in thickness.

The spilite flows are invariably fine-grained lavas and are commonly finely porphyritic with phenocrysts of feldspar from 1 to 3 mm. long, usually idiomorphic. On smooth-weathered and water-polished surfaces, a plexus of minute feldspar laths in the groundmass as in the diabasic structure is clearly seen. Amygdaloidal flows are common; flowage structure is infrequently seen on water-polished surfaces.

All of the spilites present a decomposed appearance, in greater or lesser degree, but a considerable variety in appearance may be distinguished by the "degree of freshness" displayed by different flows. Their prevailing color is a dark greyish green or a greyish purple, the latter color displayed by flows which have a high hematite content. The most highly decomposed flows are typical greenstones—dark, greyish-green rocks, usually massive or but slightly schistose, but sometimes markedly schistose. These rocks obviously have a high ferromagnesian content in the mode of chlorite and/or uralite and the numerous, small feldspar phenocrysts are largely altered to epidote. Other flows are found in which numerous fresh feldspar pheno-

crysts are found together with completely epidotized ones in a dark "greenstone" groundmass, the rock thus having a certain appearance of freshness, while it is still highly decomposed. Still other flows of dark, greenish-grey or iron-grey color have almost all of their phenocrysts composed of fresh feldspar, and under the lens numerous small points of reflection from fresh feldspar cleavages appear in the fine-grained groundmass. The latter has, however, a distinct dark green color suggestive of a high chlorite and epidote content. Epidote and carbonate masses sometimes occur in the flows and from the field relations appear to be syngenetic with respect to them. There is, therefore, a variation in the spilites from highly decomposed rocks having a low content of fresh feldspar and a high chlorite, and/or uralite, epidote, and carbonate content, through types with more abundant fresh feldspar, and a relatively lower content of epidote and carbonate. Throughout all these rocks, however, the texture, the composition of the feldspar, and the alteration products are the same. The rocks differ only in their quantity of fresh feldspar and of the alteration products, epidote and carbonate. In the flows with the highest content of fresh feldspar (albite) as in those with little fresh feldspar, no trace of the primary ferromagnesian mineral remains, and the ferromagnesian content in the mode of chlorite is equivalent to that of diabase or basalt.

Thirty-three thin sections of the spilite were examined. The rock has a characteristic texture, the groundmass consisting of a plexus of small feldspar laths, intergrown with chlorite and/or uralite, and the small phenocrysts which are usually present are of two sizes, giving a seriate porphyritic texture. The grain is constantly fine throughout all sections examined, the phenocrysts are most commonly 1.0 and 2.0 mm. long, and the groundmass laths vary between 0.07 and 0.3 mm.

The feldspar of the phenocrysts and groundmass laths, both in decomposed flows with low feldspar content and in those with a high feldspar content, is invariably fresh, and frequently displays clear albite twinning. In all sections examined its index is either less than or about equal to that of balsam. Measurements of Ng were made in index liquids and the greater number of these were below 1.540, but all measurements were less than 1.545. The extinction angle for

the phenocrysts measured on 010 is commonly 16°. For the groundmass laths the angle varied between 12° and 18° in different specimens. The feldspar, therefore, ranges from albite oligoclase roughly $Ab_{85}An_{15}$ to rather pure albite $Ab_{97}An_3$, neglecting the content of orthoclase molecule. Specimens with higher content of fresh feldspar commonly have the purest albite. In specimens with lower content of fresh feldspar, the phenocrysts and groundmass laths are largely pseudomorphed by epidote and zoisite, calcite and secondary quartz, with fresh albite forming the remainder of the altered feldspar crystal. In the rocks with intermediate content of fresh albite, in addition to the feldspar phenocrysts largely altered to epidote, zoisite and carbonate, the albite phenocrysts and groundmass laths are riddled with small epidote and zoisite grains, the latter often in short well-formed prisms arranged parallel to the albite twinning lamellae. Some phenocrysts when examined between crossed nicols have a "moth-eaten" appearance caused by the numerous small dark spots or low polarizing areas of small masses and prisms of zoisite. In the rocks with the highest content of fresh albite, these alteration products are still rather abundant and though phenocrysts largely pseudomorphed by epidote, zoisite, or carbonate, are only sparingly found, small grains of these minerals are scattered through the otherwise fresh phenocrysts and groundmass laths. The epidote, zoisite, and calcite sometimes occur in irregular random patches, but their common and characteristic occurrence is within the albite crystals. Potash feldspar has not been found in the spilites with the exception of a few cases of flows in contact with intrusive granite, and the distribution of the orthoclase in veinlets with quartz here clearly indicates contamination from the granite.

Quartz occurs in small amounts in many sections, but in some very little can be detected. It seldom appears to be in excess of the amount which would be expected as a secondary product along with epidote and other secondary minerals formed from the alteration of a more calcic plagioclase and augite. The bulk silica content of any of the spilite flows where these are unaltered by the granite is probably not different from that of a basic andesite or basalt.

The ferromagnesian content is present in the mode of secondary fibrous amphibole and/or chlorite. The secondary amphibole is a

pale, yellowish, fibrous variety with moderate pleochroism. A maximum extinction of 25° has been obtained from the better-formed crystals. It occurs as incomplete crystals with frayed edges, much of it in narrow fibrous patches between the groundmass feldspar laths, about equal to them in size, and apparently formed from a primary pyroxene. In some sections it is largely altered to biotite and chlorite,

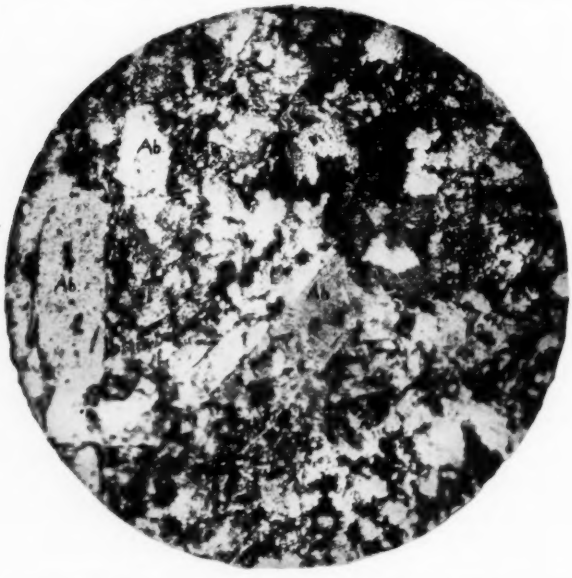

FIG. 2.—Spilite with intermediate content of fresh albite. (*Ab*) albite, containing patches of zoisite and epidote. Black areas in upper field are chlorite. Crossed nicols ×55.

the chlorite replacing the biotite. Some of this secondary amphibole is actinolite of a variety low in iron and having weak pleochroism. In the greater number of spilite sections, both in specimens with high and low feldspar content, the sole ferromagnesian mineral is chlorite. In one section chlorite was found pseudomorphic after augite with associated hematite and titanite in very small grains, and a little remnant augite. The arrangement of the latter and of the chlorite and fibrous patches of amphibole in the groundmass of other

sections indicates that the primary augite had probably a granular growth in the groundmass rather than the ophitic habit. The chlorite is confined to the groundmass, but occasionally patches of a dark, chloritic substance occupy the central part of, or form embayments into, the feldspar phenocrysts. These patches are probably altered basic glass inclusions or residuals. The chlorite occurs abundantly

FIG. 3.—Spilite, same field as above, plane polarized light ×65. (*Ep*) epidote (the larger subhedral crystals pseudomorphic after feldspar phenocrysts). (*Ch*) chlorite. Black areas are ilmenite-leucoxene aggregates.

in the flows with high fresh albite content moulded "pseudo-ophitically" about the albite laths of the groundmass, completely filling the spaces between them. It has weak pleochroism and dark Berlin blue interference colors and is probably penninite or prochlorite. The groundmass appears quite dark between crossed nicols owing to the abundant chlorite, and the plexus of albite laths stands out clearly against this background. No olivene or pseudomorph of olivene has been found in the spilite.

In the hematitic spilite, the hematite occurs abundantly in small grains in the groundmass interstitial to the albite laths. In one section the hematite appears to be pseudomorphic after pyroxene and it is possible that it is secondary after augite. This apparently would require a removal of some magnesia, however, since it has been found that specimens with a high hematite content have a relatively low

FIG. 4.—Spilite, with high content of fresh albite. (*Ab*) albite. A corner of an albite phenocryst containing an area of epidote appears at the bottom center of field. Dark areas are chiefly of chlorite and ilmenite-leucoxene aggregates. Crossed nicols ×55.

chlorite content. The hematite does not appear to pseudomorph magnetite.

Ilmenite commonly occurs in moderate amount in the groundmass, interstitial to the feldspar and largely altered to leucoxene. Large subhedral crystals of ilmenite are occasionally seen, altered to an aggregate of leucoxene with remnant patches of brownish ilmenite, or to aggregates of leucoxene and chlorite. Titanite in distinct grains is a common accessory and in some sections it is abundant, oc-

curring in fine granular form in the groundmass. A little apatite in small prisms is commonly present.

An incipient variolitic structure is occasionally developed. Radiate bunches of the groundmass albite laths may be seen here and there, and sometimes the greater part of the groundmass has this structure.

FIG. 5.—Spilite, showing lath-shaped habit of groundmass albite. Dark areas are chiefly chlorite and hematite. Crossed nicols ×55.

The spilite on the whole has not been severely sheared, and in the greater number of sections the characteristic primary structure of the groundmass is but little disturbed. No pillow structure has been recognized in any of the flows.

The altered basalt flows, found intercalated with the spilite flows in one of the greenstone groups, are similar in hand specimens to the most altered variety of spilite. In one thin section of this rock remnant acid labradorite having the composition $Ab_{48}An_{52}$ is present, but the feldspar has been largely altered to epidote, carbonate, and

fine mica (paragonite?). In other sections practically no feldspar remains, and aggregates of epidote and carbonate or of epidote and fine mica sometimes preserve the euhedral form of the feldspar. The ferromagnesian minerals are similar to those of the spilite—secondary, fibrous amphibole and chlorite. Chloritic aggregates or altered basic glass residuals are common. In these flows, with remnant lab-

FIG. 6.—Keratophyre. Broken phenocrysts of albite (*Ab*) in cryptocrystalline groundmass, containing some secondary quartz (*Q*) and a spherulitic growth (in circle). Crossed nicols ×15.

radorite, or with their feldspar almost completely destroyed, secondary albite has not developed. Furthermore a fine micaceous (paragonite?) alteration is conspicuous in these flows, whereas in the spilite flows, namely those in which secondary albite has developed, mica is not found among the alteration products of the primary feldspar. The flows with completely destroyed feldspar and without secondary albite are, therefore, classed with the similarly decomposed flows

which contain a little remnant labradorite, as altered basalt. The fine mica, if it be paragonite, may have been derived from the albite fraction of primary labradorite. The difference in alteration found in the altered basalt and spilite is believed to be due primarily to an original difference in composition between the basaltic and spilitic lavas, as discussed on following pages.

KERATOPHYRE AND QUARTZ KERATOPHYRE

The quartz keratophyre is an extremely fine-grained or felsitic rock with a cherty appearance when light or buff colored, and with a flinty appearance when dark colored. Its color ranges from buff through various shades of light grey with a faint greenish tint, to dark, iron grey and black. A pronounced greenish coloration, characteristic of all other volcanic and pyroclastic rocks of the area, is absent from the keratophyric flows, due to their relatively low content of chlorite and epidote. The black variety, like the light-colored specimens, weathers to a white or buff surface. In some of the lighter colored flows a marked flowage banding with cream and faintly greenish-colored bands may be seen.

The keratophyre flows are usually highly porphyritic with very numerous feldspar phenocrysts commonly 5 mm. or more in length. The quartz keratophyre contains less numerous and somewhat smaller feldspar phenocrysts and in addition numerous phenocrysts of glassy quartz from one to several millimeters.

The microscope shows that the groundmass of the keratophyres is cryptocrystalline to microcrystalline with rather numerous phenocrysts of albite and quartz. The larger feldspar phenocrysts are commonly euhedral, but in addition very numerous, broken fragments of quartz and albite phenocrysts exist which have been broken down by primary flowage movements or autoclasis.

In the unbroken albite phenocrysts, clear albite twinning is frequently seen, but in the broken fragments the twinning is secondary. Some resorption of the phenocrysts by the ground-mass has occurred. The feldspar of the keratophyric flows usually carries only a moderate micaceous (paragonite?) alteration and it is often remarkably fresh with only a very faint dusty growth of fine mica. It is invariably al-

bite with a composition varying between Ab_{90} and Ab_{95}. Determinative tests for potash feldspar by staining methods made on polished sections[2] of the keratophyres were negative.

In plane polarized light the groundmass sometimes has the pale, brownish color of cryptocrystalline devitrified material. It is quite free from fine chlorite. Incipient spherulitic structure may occasionally be seen, but no perlitic cracks or other features indicating devitrification were noted.

No primary ferromagnesian mineral has been found in the keratophyric flows. Pseudomorphs of epidote after an amphibole, however, are not uncommon. The subhedral form of the amphibole crystal is usually well preserved and the 124° angle between the 110 faces is frequently seen. Epidote with some zoisite makes up the bulk of the pseudomorphs, with associated carbonate, chlorite, and small titanite-leucoxene aggregates and sometimes a little hematite. In some sections, epidote pseudomorphs and aggregates of secondary pale yellow hornblende, serpentine, and actinolite are alteration products of former pyroxene phenocrysts. The original ferromagnesian minerals were clearly calcic amphibole or pyroxene rather than sodic varieties, considering the composition of their alteration products. On this account, the flows are classed as quartz keratophyres rather than as soda rhyolites, following Wells.[3]

Leucoxene masses with euhedral to subhedral outline are apparently pseudomorphs of ilmenite. Hematite in minute grains is abundant in some flows, and a little magnetite in small octahedra and minute prisms of apatite are occasional accessories.

SPILITIC TUFFS AND TUFFACEOUS SEDIMENTS

Spilitic tuffs and tuffaceous sediments, the latter predominating, make up a large proportion of the bedded succession of Caledonia massif. All gradations apparently exist between the purely pyroclastic rocks and the fine-grained tuffaceous sediments which are mixtures of pyroclastic and detrital material. The purely pyroclastic

[2] A. Gabriel and E. P. Cox, "A Staining Method for the Quantitative Determination of Certain Rock Minerals," *Amer. Miner.*, Vol. XIV (1929), pp. 290–92.

[3] A. K. Wells, "The Nomenclature of the Spilitic Suite. Part I: The Keratophyric Rocks," *Geol. Mag.* (1922), p. 352.

rocks are of chief interest, since they have a mineral composition identical with that of the spilite flows and yet are clearly of pyroclastic origin. They are very fine-grained, dull-green rocks, usually of a very dark, greyish-green color, but occasionally of a medium to light greenish-grey shade. In rather curious contrast to their dark green color and obviously high chlorite content, they weather with whitish feldspathic selvages. Only specimens of massive or but slightly sheared and schistose beds are here described, since little may be learned from thin sections of the more schistose beds which are chlorite schists consisting of recrystallized chlorite albite and secondary quartz.

In the more massive and fine-grained specimens, no texture is discernible, and even under the lens minute fragments of feldspar may be seen only occasionally. The darker specimens resemble very closely the very fine-grained, chilled borders of basic dikes. A fine bedding lamination, from one to several millimeters in width, is more frequently seen in these rocks than in the tuffaceous sediments. All gradations exist between the very fine-grained tuffs and ashy or lithic tuffs in which numerous small fragments of dark green or purple spilite, crystal fragments of albite, and dark-green chloritic spots appear in hand specimens. The fragments of spilite vary between $\frac{1}{8}$ inch and 1 inch in length and, in some beds of fine breccia, attain a length of several inches.

In thin section, the spilitic tuff is a very fine-grained rock, usually microcrystalline to cryptocrystalline. A grain of about 0.006 mm. is common, measured on angular grains of feldspar in the matrix, in which occur also larger fragments of albite which vary between .02 and .03 mm. in some beds, up to 0.5 mm. in others. The cryptocrystalline matrix in some specimens appears to be secondary chalcedonic silica, probably deposited from infiltrating ground waters in the consolidating tuff bed. Occasionally a patch of more coarsely crystalline secondary quartz may be seen. Some sections consist wholly of an even-grained matrix, but most of them contain the larger albite fragments noted. Very fine chlorite and very fine granular epidote and epidote dust are abundant in the matrix, and minute grains of titanite may sometimes be distinguished. Small cloudy patches and streaks of leucoxene sometimes containing small grains of ilmenite

occur here and there in the matrix lying parallel to the bedding lamellae. Minute grains of hematite are commonly present, and in some specimens are quite abundant. A faint vitroclastic structure is sometimes seen in the matrix when the section is examined with moderate magnification with the condensing lens thrown out and the tube raised so that the section is very slightly out of focus, in the manner suggested by Pirsson.[4] Shading with the iris diaphragm below the stage helps to bring out the structure more clearly. The vitroclastic structure is very faint and could not be photographed, or drawn on a camera lucida, but in some sections it is unmistakable. It consists of small cusp-shaped and hook-shaped forms in the matrix, which are best seen in the sections with secondary chalcedonic silica in the matrix, and are enhanced sometimes also by the fine coating of epidote dust, although the fine epidote and leucoxene usually mask the structure. The vitroclastic forms presumably represent devitrified or silicified fragments of spilitic glass, the forms very faintly preserved by the very fine-grained silicification or outlined by the growth of very fine epidote dust as alteration material within fragments of feldspathic or ferromagnesian composition. The structure, although less clear, is entirely similar to that found by the writer in rhyolitic tuff beds of Carboniferous age at Marblehead Neck, Marblehead, Massachusetts.

The larger crystal fragments of albite which occur embedded in the matrix have an index less than that of balsam, are optically positive, and very often display clear albite twinning. They contain no inclusions of the matrix and are frequently found clear of the epidote, calcite, or quartz which are so constant in the phenocrysts of the spilite flows. They sometimes contain small patches of a chloritic substance which probably represent altered glass inclusions. Occasionally they have subhedral forms, but they usually have a partly jagged border indicating that the fragments are broken euhedral or subhedral forms. The albite fragments are not phenoclasts produced by shearing or by flowage. No shearing or flowage structures are found in the sections of the more massive beds here described. The albites are therefore pyroclastic crystal fragments. Irregularly

[4] L. V. Pirsson, "The Microscopic Characters of Volcanic Tuffs," *Am. Jour. Sci.*, Vol. XL (1915), pp. 191–211.

shaped patches of fine chloritic material which are occasionally seen as units in the matrix, sometimes adherent to albite fragments, probably represent altered basic glass.

The spilitic tuffs have the same mineral composition as the spilite flows: albite, very little quartz—except where silicified—epidote and chlorite, titanite and leucoxene, hematite and probably a little ilmenite. The lithic tuffs are composed largely of fragments of spilite and albite crystal fragments, with an aggregate of the above minerals forming the matrix. The purer tuff beds, considering their fine bedding laminae and even grain and their association with conglomeratic and other partly detrital sediments, are probably water-laid tuffs.

GENERAL CHARACTER AND PROBLEMS OF THE SPILITIC ROCKS

The problem of the spilitic rocks has been discussed most extensively in the petrographical literature by the British authors, Sir J. J. H. Teall,[5] E. B. Bailey and G. W. Grabham,[6] and Sir J. S. Flett and H. Dewey,[7] through whose papers the problem first attained prominence. The name "spilite" was given first by Brongniart[8] to flow rocks exhibiting the pillow structure, and was used first in English in this sense by Bonney.[9] Later Termier[10] discussed the rocks from Mt. Pelvoux in which an alteration somewhat similar to that in spilites had occurred.

Spilitic rocks occur in great abundance among the Devonian schalsteins of Germany and Bohemia and an extensive literature on these rocks is extant,[11] but the problem of the spilites understood by the present significance of the term in petrographical literature apparently is not dealt with in these writings. The term "spilite," orig-

[5] *Mem. Geol. Surv.* (1899), p. 85.

[6] "Albitization of Basic Plagioclase Felspars," *Geol. Mag.* (1909), pp. 250–56.

[7] "On Some British Pillow-Lavas and the Rocks Associated with Them," *Geol. Mag.* (1911), pp. 202–9 and 241–47.

[8] Al. Brongniart, *Class. et caract. min. des roches* (1827), p. 98.

[9] T. G. Bonney, "On the So-called Spilites of Jersey," *Geol. Mag.* (1893), p. 59.

[10] *Bull. Soc. Geol. Fr.*, Vol. XXVI (1898), p. 165.

[11] References to the extensive literature of the Devonian igneous rocks of Germany and Bohemia may be found in H. Rosenbusch, *Mik. Physiographie*, and in F. Zirkel, *Petrographie*.

inally used in a textural sense, has, owing to the work of the British authors previously mentioned, come into general usage in a chemical-petrographical sense, as designating femic rocks with a high soda-content and containing an acid plagioclase as the dominating feldspar. A highly decomposed state for the rocks is also generally considered to be implied by the term, but according to Benson, Wells, and Sundius,[12] this is not general or necessary for the rocks belonging to the spilite group. An interesting occurrence of spilitic rocks has been described by Benson from New South Wales, Australia,[13] where, in the Tamworth district, a continuous series of keratophyric, spilitic, and doleritic lavas occur, the minerals of which are primary and show few signs of alteration.

The spilite problem has been discussed further by Wells,[14] and in a recent paper Sundius[15] discusses the chemical and petrographical relations of the spilite group proper, and the relations of spilites to other magmatic rocks of equal acidity.

The texture of the British spilite flows described by Dewey and Flett varies but slightly; frequently they appear to have contained a fair amount of glassy base, though sometimes they consist wholly of feldspar laths with fluidal arrangement. Their feldspars are occasionally microporphyritic and some have large feldspar phenocrysts. A large number of spilites have the variolitic structure. With regard to the mineral composition of the British spilites as described by Dewey and Flett[16] and by Wells,[17] the feldspar is always a sodic plagioclase ranging from pure albite to albite-oligoclase, except where some remnant basic plagioclase is present. The proportion of potash shown in the analyses is always very small. The primary ferromagnesian mineral is invariably altered to chlorite (and to uralite in the Kiruna spilites described by Sundius[18]) and, in all cases

[12] Publications cited in following footnotes.

[13] W. N. Benson, "Spilitic Lavas and Radiolarian Rocks in New South Wales," *Geol. Mag.* (1913), pp. 17–21.

[14] *Op. cit.*, pp. 346–54; and "Part II: The Problem of the Spilites," *op. cit.* (1923), pp. 62–74.

[15] N. Sundius, "On the Spilitic Rocks," *Geol. Mag.* (1930), pp. 1–17.

[16] *Op. cit.*, pp. 203–4.

[17] *Op. cit.*, "The Problem of the Spilites," p. 71.

[18] *Op. cit.*, p. 3.

where the original ferromagnesian mineral remains, it is a brownish colored augite, usually granular, but occasionally subophitically intergrown with the feldspar. Remains or pseudomorphs of olivene have sometimes been discovered by careful search, but its presence is rarely recorded in spilites.

Though the majority of spilites described are altered rocks in that the abundant, original ferromagnesian mineral has been largely or completely altered to chlorite or uralitic hornblende and an original and more anorthic plagioclase has been pseudomorphed by fresh albite with accompanying alteration products, epidote, zoisite and/or clinozoisite and calcite, there are found, among the spilites, rocks which are apparently little altered, the feldspar presumably primary and the ferromagnesian minerals but little, or not at all, changed. Spilites of this kind are found in the Tamworth District, New South Wales, described by Benson,[19] the feldspar of which is dominantly albite or albite-oligoclase with a composition An_7–An_{15} with oligoclase and in rarer cases andesine occurring as kernels. The analyses of the Australian spilites agree closely with those of the decomposed spilites of Cornwall, England. The spilites of the southernmost part of the Kiruna greenstone zone described by Sundius also are relatively but little altered. The spilitic rocks therefore may not be considered merely as altered lavas and disposed of as "altered andesites" etc., nor may the more altered types be ignored, and only the little altered or unaltered types may be considered as true spilites. The latter appears evident when one considers that analyses of the unaltered Australian spilites agree closely with those of the generally altered British Devonian spilites, and that the soda-rich, keratophyric, salic rocks, characteristically associated with the spilites, are as prominent in volcanic piles in which the spilite is of the altered type as in those in which the unaltered type prevails.

The problem of the spilites, therefore, relates to the cause and time of the alteration of the rock, the source of the altering agents, and the relation of these agents and the bulk composition of the altered rock to the original magma. With the latter question is associated that of the primary plagioclase, whether this be sodic or calcic, and of the existence of a separate suite of rock magmas.

[19] *Op. cit.*

The process of replacement of more basic plagioclase by albite with the formation of the attendant alteration products, epidote, zoisite, and calcite, is known to take place in several ways. It is sometimes the result of weathering or it may arise through shearing or dynamic metamorphism. Further, albite may replace lime-soda feldspar by a process of albitization which may be a juvenile, post-volcanic pneumatolytic mode of alteration,[20] essentially a deuteric process, or finally by a process of hydrothermal alteration, specifically that of wallrock alteration accompanying vein or lode formation.

Dewey and Flett[21] have shown clearly that the secondary albite in spilites has not formed by the process of weathering or through dynamic metamorphism. The opinion of the British authors was that the spilitic rocks, in those cases where they are rich in decomposition products, were altered by an autometamorphic change, soda and carbon dioxide being retained in solution during the solidification of the magma, after which they acted upon the minerals of the rock. The calcic minerals would be decomposed, the CaO content at the same time being fixed in epidote and calcite or possibly carried away. The keratophyres and quartz keratophyres, albite diabases, dolerites, and soda granites found associated with the British spilites were included by Dewey and Flett in a special petrographical group, the "spilitic suite."

Sundius[22] in his recent paper maintains that the spilites are distinguished petrographically as a separate rock type. The chemical characteristics of the spilites are high content of soda, ferrous iron, and of TiO_2, and very little potash, and a deficiency of alumina as compared with that of equally femic rocks such as diabase and basalt. In the Kiruna spilites, Sundius finds a considerable variation in the quantity of secondary products present, some flows being rich in epidote and calcite, though others are relatively poor in these minerals and correspondingly richer in pure albite. This variation he believes to be due in part to a movement of the epidote within the rocks during their alteration, so that some parts would have a concentration of epidote but other portions would be poorer in it and relatively

[20] E. B. Bailey and G. W. Grabham, "Albitization of Basic Plagioclase Felspars," *Geol. Mag.* (1909), p. 250.

[21] *Op. cit.*, pp. 204–5.

[22] *Op. cit.*, p. 14.

richer in fresh albite; and in part to original differences in the feldspar of the rocks, the spilites with the smaller quantity of epidote and calcite having had a more sodic primary feldspar than those with more abundant decomposition products. It is pointed out, however, that there is little ground for the hypothesis of a substantial change in the CaO–Na_2O ratio of the rocks through their mineral alteration, except locally owing to an uneven distribution in the rocks of the alteration products, and that the bulk composition of the rocks has remained unchanged.

The feldspar mixtures calculated from the analyses of spilites vary within rather wide limits, the normative feldspar containing between 7 and 30 per cent An. for the spilites proper. It is a well-known fact that the normative feldspar calculated for spilites does not agree either with the feldspar of the mode (albite) or with what might be supposed to have existed in the unaltered rock (basic plagioclase), considering the otherwise basic nature of the rocks. In the Karelian spilites in which the pyroxene is undecomposed, Sundius[23] has calculated a normative plagioclase with 14 per cent anorthite, part of which has entered the pyroxene. After allowing for the clinozoisite of the mode, which is calculated as an alteration product of former anorthite in the plagioclase in which it is now distributed, he obtains a plagioclase containing about 10 per cent anorthite. This he considers to be the approximate primary feldspar of these rocks, though for spilites in general the same author concludes that any moditive plagioclase containing more than 5 per cent anorthite may be considered as a remnant of the original feldspar formed during consolidation of the magma. Sundius[24] maintains that to assume an original basic plagioclase in such rocks as the relatively unaltered spilites of New South Wales and the epidote-calcite poor spilites of the southern Kiruna district requires that a real alteration in the bulk composition of the rocks has occurred, an alteration for which there is no evidence.

The New Brunswick spilites agree closely with the general character and composition of the various spilites reviewed on foregoing pages. They are rocks characterized by a sodic plagioclase and a high content of femic minerals; the titanium minerals, titanite, and

[23] *Ibid.*, p. 15. [24] *Ibid.*

leucoxene (with ilmenite) are abundant and constantly present, and the absence of potash feldspar is evident both from microscopic study and from determinative staining tests. Sundius' views on the spilitic rocks have been presented rather fully by the present writer, since the New Brunswick spilites have been found to have many features in common with those of the Kiruna greenstone belt. The rocks are lava flows, although apparently lacking the pillow-structure, interbedded with tuffaceous rocks and accompanied by salic keratophyric flows. Some movement of the epidote was found to have occurred in certain flows, as in the Kiruna rocks, with parts of flows relatively enriched in epidote occurring in irregular patches and in discontinuous veinlets, which from field relations are apparently syngenetic with respect to the flows. Further, there occur among the spilites types which are rich in fresh albite and types which have a lower content of fresh albite and a high content of the decomposition products, epidote, zoisite and calcite; and with the spilites there are associated highly decomposed flows, with remnant basic plagioclase, or with wholly decomposed feldspar or but very little secondary albite, which must be regarded as altered basalts.

A considerable variation is found in the amount of the secondary minerals, epidote, zoisite, and calcite in the rocks, as in the Kiruna spilites. This is probably in part due to a movement of the epidote (and calcite) within the flow, as previously mentioned, and in part to original differences in the primary feldspar of the rocks. In the types with a very high fresh albite content, the idiomorphic albite phenocrysts and groundmass laths have all the appearances of primary magmatic crystals and though the amount of granular epidote, zoisite, and calcite distributed through the albite clearly indicates somewhat more anorthic primary plagioclase, the latter may not be assumed to have been a basic one in these rocks, without admitting that an alteration of the bulk composition of the rock and of the CaO–Na_2O ratio has occurred. At the most, the primary feldspar of these Ab-rich types could not contain more than 15 per cent anorthite. The types with lower content of fresh albite and higher epidote-calcite content contained a more anorthic primary plagioclase with probably 20–30 per cent anorthite.

The mineral alteration in the New Brunswick spilites has not been

caused by shearing of the rocks. The spilites as a group are rather massive, dense green rocks, and the flows with the highest content of fresh secondary albite are the least sheared and are entirely massive. The albite does not occur as recrystallized porphyroblasts, but has pseudomorphed the phenocrysts and diabasic groundmass laths of the primary feldspar, and where the rocks are sheared, the albite crystals are broken or bent and show secondary twinning.

The albitization of the spilites is not due to an introduction of soda to the rocks from an outside source. The fact that albite granite is intimately intruded into the bedded rocks of the area suggests that regional contact metasomatism of the bedded rocks by the intruding granite may have resulted in widespread albitization of the rocks. The alteration of the spilite and of other members of the bedded rocks near granite contacts does not substantiate such a hypothesis, orthoclase having been introduced near the contacts more frequently than albite, and the andesine of the quartz diorite which has been intruded by the granite has not been albitized. Moreover, pebbles and fragments of spilite occur in quartzite conglomerate which forms a belt lying conformably within the bedded succession of flows and pyroclastics, the albitization of the spilite clearly having been completed before the deposition of the pebbles and pyroclastic fragments of this rock in the conglomerate. The conglomerate with other members of the bedded succession was folded and later intruded by the albite granite. Since the quartzite conglomerate is contemporaneous with the volcanic series, the albitization of the spilite flows must have occurred very shortly after their consolidation. In this connection, the spilitic tuffs and lithic tuffs offer additional evidence of the early formation of the albite. The lithic tuffs contain fragments of spilite from $\frac{1}{8}$ inch to 1 inch in length and the occasional beds of breccia character contain flat fragments up to 3 inches in length. These fragments contain a high proportion of femic mineral (chlorite) and the feldspar of the groundmass laths and of the occasional small phenocrysts is albite. The fragments apparently have been torn from solidified spilite flows adjacent to the explosive vent, and, since contemporaneous spilite flows are interbedded with the tuffs, the time of albitization of the spilite mass from which the lithic fragments were derived is fixed within rather

narrow limits. The fragments also may have been derived in part from a solidified plug or cap in the neck of the volcanic vent, and the albitization therefore completed in the solidified cap before the extrusion of the fragmental material from the vent. The finer spilitic tuffs and the tuffaceous sediments are composed of pyroclastic albite crystals, epidotic dust and chloritic material derived from original pyroxenic dust, and glass fragments, with granular titanite and leucoxene. The pyroclastic crystal fragments in tuffs may originate in the magma itself before its explosive eruption, or they may become disrupted from the walls of the volcanic vent or from the shattered plug formed at the mouth of the vent. Many of the angular pyroclastic albite crystal fragments in the spilitic tuff and tuffaceous sediments may be magmatic crystals, formed in the volcanic conduit immediately before explosive eruption. If the albite fragments are assumed to represent entirely broken phenocrysts of shattered spilite rock, it must be granted that a remarkably clean separation of the phenocryst fragments from the groundmass has been effected in all cases, since crystals with adherent diabasic groundmass have not been found, with the exception of the larger lithic fragments in the lithic beds. The albite of the spilitic tuffs therefore may be primary and have crystallized from a spilitic magma prior to its explosive extrusion.

The great abundance of pyroclastic rocks in this volcanic series and the basic nature of the tuffs (no salic tuffs of keratophyric composition occur) appears abnormal if the spilitic flows and tuffs are assumed to have been of normal basaltic composition at the time of their extrusion. If, however, a primary acidic plagioclase in the spilite flows is admitted, the nature and the abundance of the pyroclastic material are comprehensible, since the crystallization of acidic plagioclase or albite is rather to be expected in an otherwise basic magma which has been explosively extruded as volcanic dust and ash.

The mineral alteration in the spilite flows appears to have taken place very shortly after their extrusion and probably during the final stages of their consolidation. It seems most probable to the writer that the conversion of the original plagioclase to a more highly sodic one, or albite, was caused by a residuum of sodium metasilicate and

quartz in the lavas as postulated by Eskola. By this alteration, an original acidic plagioclase containing about 15–30 per cent anorthite was converted to albite with 2 to 5 per cent anorthite, neglecting the orthoclase molecule, and accompanied by alteration products, epidote, zoisite, and calcite. In the altered basalts, namely, the flows with remnants of primary labradorite or with their primary feldspar almost completely altered to epidote, calcite, secondary quartz, and fine mica (paragonite?), and with very little albite, soda-rich residual solutions appear to have been absent. In these flows, alteration of the primary minerals probably was effected by the water of the magma, held in glassy residuals in the rock, which subsequently decomposed the glass to chloritic aggregates and acted upon the minerals of the rock. The dark chloritic aggregates in the feldspar of these rocks are undoubtedly altered glass inclusions. Alteration of the primary ferromagnesian minerals in the quartz keratophyres apparently has been effected in the same manner, by magmatic water held in the cryptocrystalline base, though the primary albite has been unaffected. The magmatic water was apparently not rich in soda, but was probably high in CO_2 content.

The albite in the spilitic flows and pyroclastics therefore would be related either directly or indirectly to the extruded magma, and a high soda content was undoubtedly inherent in this magma. To the writer, the close association of the salic and soda-rich keratophyre and quartz keratophyre flows with the basic spilitic rocks is most significant. These flows occur, alternating with spilite flows, spilitic tuffs, and tuffaceous sediments throughout an apparently conformable, bedded succession which has a total thickness of roughly 18,000 feet. Within this succession, groups made up chiefly of spilite and quartz keratophyre flows occur, these rocks occurring interbedded as individual flows a few feet in thickness and in belts several hundreds of feet thick. This tremendous amount of extrusive material, the product of one period of volcanic activity, is characterized throughout by a richness in soda and there can be little doubt that the keratophyres and quartz keratophyres have been differentiated from the same magma source as the spilite flows. Wells,[25] describing

[25] "The Nomenclature of the Spilitic Suite. Part I. The Keratophyric Rocks," *Geol. Mag.* (1922), p. 351.

the keratophyric rocks states that "the constant association of keratophyres with typical spilites is a proven fact, and that the two are consanguineous can no longer be questioned."

Of the American occurrences of albite-rich greenstones or basic lavas, the extrusive greenstones of the Mariposa formation in the Mother Lode belt of California, described by Knopf,[26] appear among others to be most clearly of spilitic composition. The author, however, does not apply the name "spilite" to these rocks, but under the heading "Keratophyric Rocks," describes a variety of albitic volcanic rocks. Knopf concludes that the albite of the melanocratic flows, though resembling normal pyrogenetic crystals, is the result of a very perfect pseudomorphic replacement after calcic plagioclase, probably labradorite.

Though in the Mother Lode belt, and in places within the Canadian Shield, albitization plays a part in the process of wall rock alteration accompanying gold-quartz deposition, and the albitic greenstones within the mineralized zone probably owe their albite in whole or in part to such hydrothermal alteration, the study of such areas might well include an investigation into the possible spilitic character of the greenstones, if the albitic character of the latter proves to be a regional feature.

[26] Adolph Knopf, "The Mother Lode System of California," *U.S. Geol. Surv. Prof. Paper 157* (1929), pp. 16–18.

GEOLOGY OF THE MOUNTAIN UPLIFT TRANSECTED BY THE SHOSHONE CANYON, WYOMING[1]

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ABSTRACT

The region of Shoshone Canyon is significant, because it contains both overthrust and block-faulted structures in close association. The stratigraphic section comprises 3,200 feet of conformable, marine, Paleozoic sediments, 800 feet of Permo-Triassic red beds and over 5,000 feet of Jurassic and Cretaceous sandstones and shales. This is notably thinner than the corresponding section in the overthrust belt of southeastern Idaho, but greater than that in the front ranges of Wyoming and Colorado.

Flexures, which appear as folds in the Paleozoic and Mesozoic strata, have been traced with certainty to sharply faulted blocks of the underlying pre-Cambrian complex. Distinct from these is the Heart Mountain overthrust which places Paleozoic limestones above sediments of Mesozoic and early Tertiary age. The two types of structure are shown to differ in age and in the thickness of rocks involved. Upfaulting of the pre-Cambrian with flexing of the sediments took place in late Upper Cretaceous. These uplifts were extensively eroded and covered with sands and gravels before the overthrusting in late Eocene or Oligocene. Further faulting in Miocene and later time accentuated the earlier uplifts and disposed the remnants of the overthrust sheet at four different elevations.

INTRODUCTION

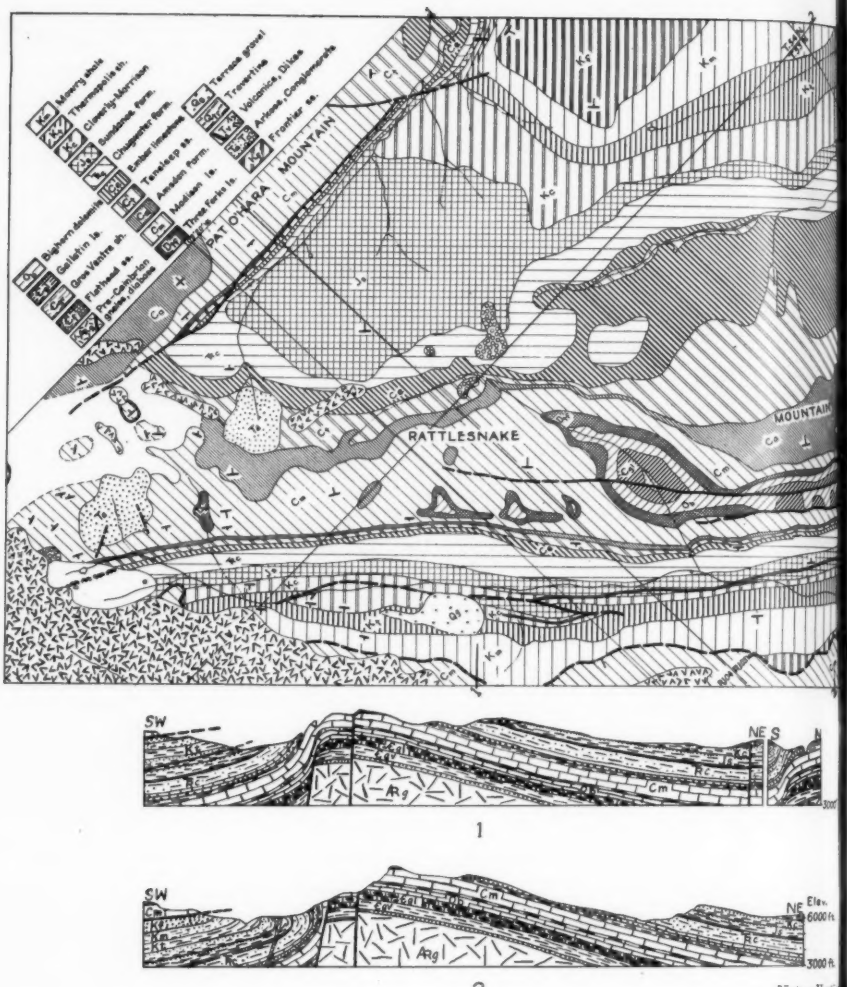
West of the Bighorn Basin in northwestern Wyoming lie the mass of volcanic flows and agglomerates of the Absaroka Range and the southern tip of the uplifted pre-Cambrian block of the Beartooth Mountains. At the margin of the Basin, as a southward physiographic continuation of the Beartooth front, is a series of mountain ridges in which pre-Cretaceous strata are uplifted (Fig. 2). The longest and southernmost of these comprises Rattlesnake and Cedar mountains, between which the Shoshone River has cut a canyon 3,000 feet deep, just west of the town of Cody (Fig. 3).

With the exception of a rough reconnaissance by T. B. Comstock in 1873, Eldridge,[2] of the U.S. Geological Survey, was the first to describe the geology of the region. Not until the work of Dake[3] were

[1] Doctor's Dissertation presented at Johns Hopkins University, February, 1934.

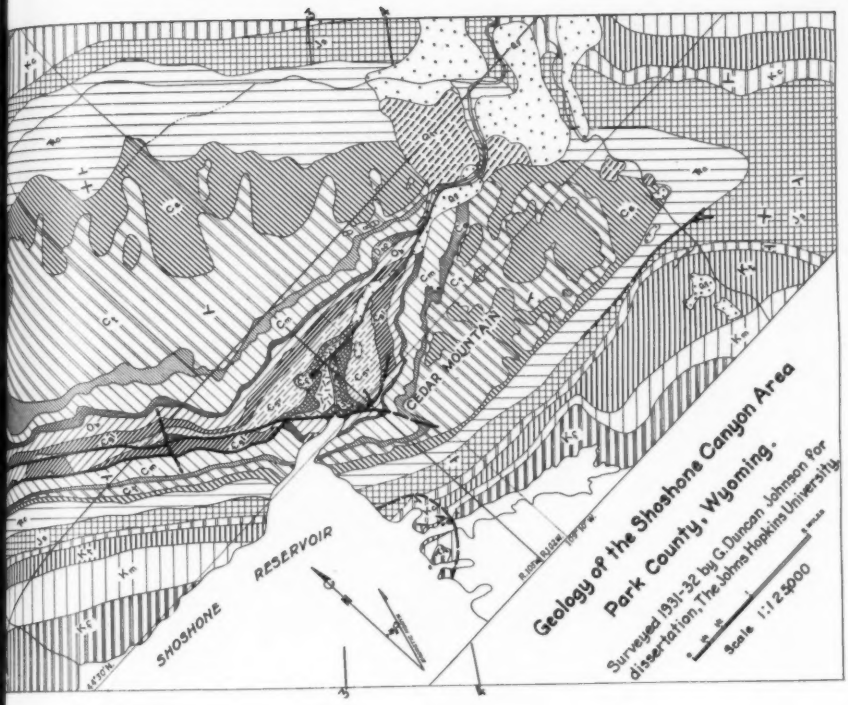
[2] George H. Eldridge, "A Geological Reconnaissance of Northwest Wyoming," *U.S. Geol. Surv., Bull. 119* (1894), pp. 29–32.

[3] C. L. Dake, "The Heart Mountain Overthrust and Associated Structures in Park County, Wyoming," *Jour. Geol.*, Vol. XXVI (1918), pp. 45–55.



Note.–Vertical scales same

FIG. 1.—Geologic structure se


Geology of the Shoshone Canyon Area
Park County, Wyoming.
Surveyed 1931-32 by G. Duncan Johnson for dissertation, The Johns Hopkins University.
Scale 1:12,000
CEDAR MOUNTAIN
SHOSHONE RESERVOIR

3

4

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structure sections

the segments of the Heart Mountain overthrust correctly interpreted. There still remained the difficulty in explaining the relation of

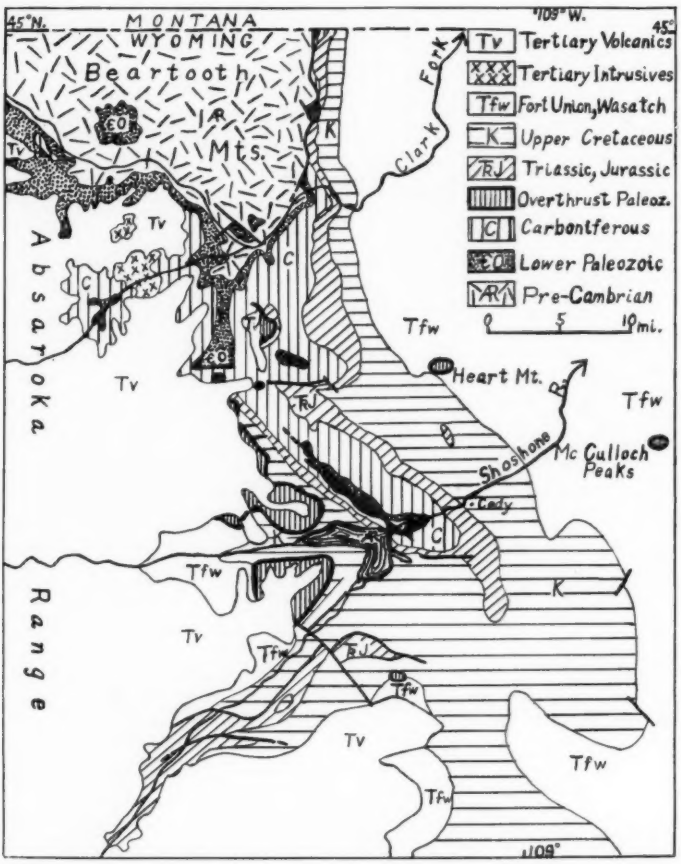

FIG. 2.—Regional geologic map to show relation of Heart Mountain overthrust and uplifts (geology of Beartooth Mountain front from R. V. Hughes. *Proc. Nat. Acad. Sci.*, Vol. XIX [1933], Fig. 1).

the "asymmetric anticline" of Rattlesnake Mountain to the overthrust. It was toward a solution of this problem that the present investigation was undertaken.

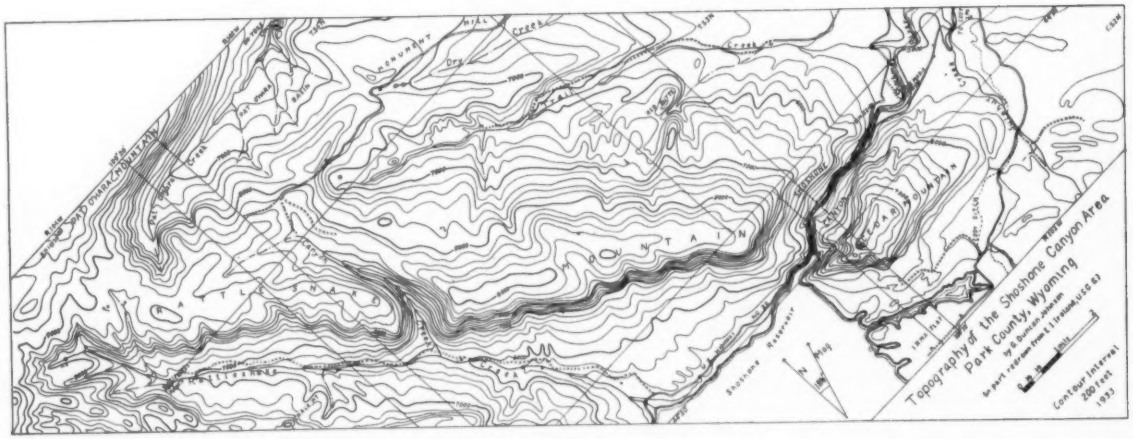
Topography of the Shoshone Canyon Area
Park County, Wyoming
Contour Interval
200 Feet
1933

FIG. 3

Field studies of Rattlesnake–Cedar Mountain, together with the adjacent valleys and part of Pat O'Hara Mountain, were carried out in August and September, 1931, and June to September, 1932. The Princeton University Geological Research Base, under the direction of Professor W. Taylor Thom, Jr., facilitated the work by the loan of field equipment and by putting the writer in close touch with other geologists working in the region.[4]

STRATIGRAPHY AND PETROLOGY

The stratigraphic section exposed in Shoshone Canyon and in the river trench below the Canyon represents an essentially conformable though discontinuous sequence from the Middle Cambrian through the Upper Cretaceous (Fig. 4). Overlying this series with pronounced unconformity are conglomerates, arkoses, and basic volcanics of Tertiary age, and terrace gravels and travertine probably no older than Quaternary.

The Paleozoic marine series consists roughly of 60 per cent limestones and dolomites, 25 per cent shales, and 15 per cent sandstones. Most investigators of the region agree that no rocks are present to represent the Algonkian, Lower Cambrian, Lower Ordovician, Silurian, or Lower Devonian. Formation names and boundaries have been adopted from descriptions by Darton, Tomlinson, Blackwelder, and Lovering[5] with little regard for the earlier division in the Yellowstone Park region.

Flathead sandstone (Middle Cambrian).—The Flathead sandstone lies on the nearly plane surface of the pre-Cambrian granite and out-

[4] To A. N. Dusenbury, Jr., Freeman Foote, T. J. Gillen, Jr., and Carl Broedel, the writer owes a debt of thanks for their services as field assistants. Howard Bell, Edward Heald, Willard Rhodes, Mr. and Mrs. Elmon Hall, and George Cless, residents of Park County, were especially helpful during the progress of the field work. A small contribution by the Cody Club toward field expenses is gratefully acknowledged. Professor Ernst Cloos, of Johns Hopkins University, has given much helpful advice.

[5] N. H. Darton, "Geology of the Bighorn Mountains," *U.S. Geol. Surv. Prof. Paper 51* (1906); "Geology of the Owl Creek Mountains," *59th Cong. 1st Ses. Senate Exec. Doc. 219* (1906); C. W. Tomlinson, "Middle Paleozoic Stratigraphy of the Central Rocky Mountain Region," *Jour. Geol.*, Vol. XXV (1917), pp. 112–34, 244–57, 373–94; Eliot Blackwelder, "New Geologic Formations in Western Wyoming," *Jour. Wash. Acad. Sci.*, Vol. VIII (1918), pp. 417–26; T. S. Lovering, "The New World or Cooke City Mining District, Park County, Montana," *U.S. Geol. Surv. Bull. 811(a)* (1929), pp. 17–29.

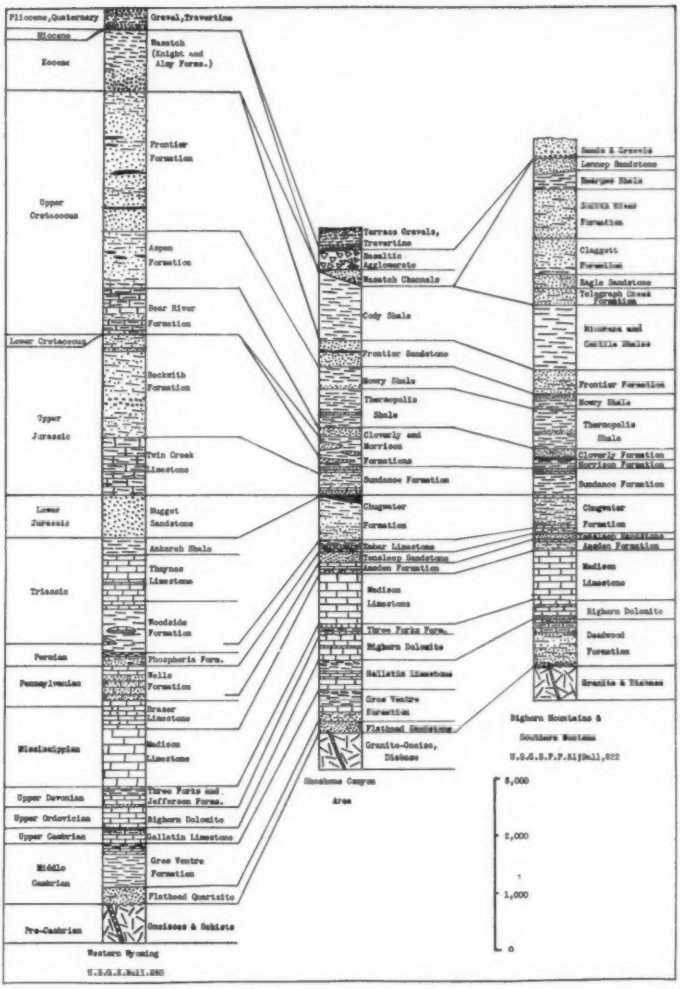

FIG. 4.—Stratigraphic column correlated with regions to the west and east

crops as a fringe along the walls of the inner gorge of Shoshone Canyon. It is a brownish, indurated sandstone or quartzite with lenses of very coarse sandstone or fine quartz conglomerate in the lower 20 feet. No persistent basal conglomerate overlies the pre-Cambrian surface. On the high bench southeast of Shoshone dam the Flathead section is:

	Feet
Sandy oölitic hematite with *"Lingulella"*	5
Medium-bedded, brown-spotted sandstone	20
Massive, medium-grained, brownish sandstone	120
	145

Gros Ventre formation (*Middle Cambrian*).—The Gros Ventre formation is composed largely of shales and thin-bedded sandstones, but certain hard beds outcrop conspicuously. The section on the south side of Shoshone Canyon is:

	Feet
Greenish limy and sandy shales; flat-pebble conglomerates	200
Hard, maroon sandstone	8
Thin-bedded sandstones and sandy shales	180
Nodular limestone with shaly beds	95
Brown, lime-cemented sandstones; limy shales	90
	573

Gallatin limestone (*Upper Cambrian*).—A massive, cliff-forming limestone, thin-bedded limestones, and flat-pebble limestone conglomerates have been grouped as the Gallatin formation. Exposures on the north side of Shoshone Canyon show:

	Feet
Flat-pebble conglomerates; shales, with trilobite fragments	200
Massive bed of limestone	12
Thin-bedded limestones; calcareous shales	120
Cliff-forming, dolomitic limestone, some nodular beds	180
	512

Flat-pebble limestone conglomerates occur in the upper Gros Ventre but are especially characteristic of the Gallatin. Ellipsoidal pebbles, 1–5 inches in diameter and $\frac{1}{4}$–1 inch thick, lie usually parallel to the bedding or occasionally at steep angles to it. A greenish-yellow calcareous shale binds the pebbles together and constitutes intervening layers free from pebbles. That these are typical intraformational

conglomerates is demonstrated by their repeated recurrence, and the presence exclusively of flat pebbles of soft limestone derived from underlying beds.

All the Cambrian beds described had been embraced in the Deadwood formation by Darton in the Bighorn Mountains. Only in regions farther west have subdivisions of the Cambrian been made. The distinctive lithologic characters, however, are sufficiently well marked to make such a division possible in the Shoshone Canyon.

Bighorn dolomite (Upper Ordovician).—All of the cliff-forming dolomites between the flat-pebble conglomerates of the Gallatin and the soft, orange limestones of the Three Forks are mapped as Bighorn dolomite. Stadia measurements of a high bluff on the north side of the Canyon west of the highway bridge give:

	Feet
Hard, silicious dolomite, usually brecciated	20
Massive, buff dolomite	125
Fine-grained, white dolomite weathering chalky	5
Massive dolomite; thin beds of sandstone	18
White, chalky dolomite	17
Massive, gray dolomite	10
White, closely-fractured dolomite	60
Light buff, massive dolomite	85
Buff and gray massive dolomite	110
	450

A fossiliferous basal sandstone (Lander sandstone) appears to be lacking in the Shoshone Canyon section. Approximately at the top of the Bighorn is a single occurrence of reddish limestone. This lenticular body yielded no fossils, but probably can be interpreted as an early Devonian filling of a channel in the Bighorn, like those described by Hughes[6] as "basal Jefferson" along the Beartooth Mountain front.

Three Forks formation (Upper Devonian).—Overlying the Bighorn with few evidences of unconformity is a group of soft, orange-buff limestones and yellow-green shales which are correlated solely on the basis of lithology with the Three Forks formation as described by

[6] R. V. Hughes, "The Geology of the Beartooth Mountain Front in Park County, Wyoming," *Proc. Nat. Acad. Sci.*, Vol. XIX (1933), pp. 239–53.

Tomlinson and Lovering.[7] Beds of shale and limestone with a few thin sandstones alternate to give a total thickness of 180 feet. Possibly the upper 40 feet of the measured Bighorn section, together with 80 feet of beds with typical Three Forks lithology, represent a diminished thickness of the Devonian Jefferson dolomite.

Madison limestone (Mississippian).—Following directly upon the Three Forks is the great thickness of Madison limestone. This is by far the most conspicuous formation in the area, forming two distinct series of cliffs on either side of Shoshone Canyon and a remarkable line of palisades above Rattlesnake Valley (Fig. 7). Wherever remnants of the Heart Mountain overthrust occur, Madison limestone forms a conspicuous cliff. The entire thickness of the Madison shows little variation in general character; such features as lenses of coral-reef limestone, brecciation, and solution cavities are irregularly distributed from bottom to top. No persistent fossil horizons could be distinguished, although the formation as a whole is noticeably fossiliferous. Because of the lack of key beds and the secondary brecciation of much of the Madison section, the measured thickness of 900 feet is at best an approximation.

Amsden formation (Pennsylvanian).—The soft limestones of the Amsden formation are usually concealed beneath a mantle of reddish soil. Only the basal quartzite is conspicuous as a capping of the highest crest of Rattlesnake and Pat O'Hara mountains. A composite section from the north side of Shoshone Canyon is:

	Feet
Blue chert	1
Cross-bedded, brown sandstone	10
Sandy limestone with chert	15
Red-purple sandstone	6
Soft, buff to pink limestone	60
Thin-bedded purplish limestone and shale	70
Cross-bedded indurated sandstone	0–10
	172

Tensleep sandstone (Pennsylvanian).—Because of its great resistance to weathering, the Tensleep sandstone occupies a conspicuous place in the areal geology. An additional area is covered by broken blocks of sandstone to such a depth that the talus might well be con-

[7] *Op. cit.*, pp. 124, 383–85; *op. cit.*, p. 28.

sidered a geologic unit in itself. The Tensleep in place comprises 180 feet of massive, usually cross-bedded sandstone, white on a fresh break but weathering to a characteristic reddish-brown. With the Tensleep the uniform marine sedimentation over a large region during the Paleozoic came to an end.

Embar limestone (Permian).—The term "Embar" has been retained to designate a series of shaly and cherty limestones which represent the extension into northern Wyoming of the Phosphoria, Dinwoody, and part of the Park City formations of western Wyoming and southeastern Idaho. The extreme variation in thickness and facies of the Embar and its equivalents is discussed by Condit,[8] who shows that the phosphatic limestone facies of the Wind River Mountains becomes thinner toward the east and intertongues with red bed facies. On the slopes of Cedar Mountain the Embar includes:

	Feet
Yellowish-calcareous shales	30
Nodular limestone	10
Red sandstone	4
Yellow-green, shaly limestone	50
Platy, sandy limestone	27
Sandy limestone with few chert nodules	30
Thin-bedded limestone (chert and lime concretions)	15
Brown, massive limestone (chert and calcite vugs)	18
Yellow, sandy limestone	16
Nodules of chert cemented by lime	10
	210

Chugwater formation (Permo-Triassic).—Red beds of the Chugwater formation are typically developed around the Rattlesnake Mountain uplift. The general succession is:

	Feet
White, saccharoidal and massive gypsum	0–90
Red shales with seams and flakes of gypsum; interbedded red sandstones toward base	410
Interbedded red sandstones and shales, with a few yellow sandstone beds	300
	800

[8] D. D. Condit, "Relation of the Embar and Chugwater Formations in Central Wyoming," *U.S. Geol. Surv. Prof. Paper 98 (o)*, (1917).

Extreme local variations in thickness, especially of the gypsum member of the Chugwater, are only in part the result of unconformities. Condit[9] has suggested that thinning may result from solution of gypsum or other salines. Along the sharp flexures of Rattlesnake and Pat O'Hara mountains the gypsum and a large part of the shales appear to have been squeezed out during deformation. Neither the "red bed conglomerates," nor the Alcova limestone, nor the fossiliferous Popo Agie facies has been found in the Cody region; consequently the Permian, the Triassic, and the Jurassic portions of the red beds cannot be distinguished.

The stratigraphic section from the base of the Sundance to the Eocene Wasatch beds has been described in sufficient detail by Hewett.[10] Only slight changes have been made in the grouping of the units described into formations.

Sundance formation (Jurassic.)—A summary of the Sundance section measured by Woodruff and Hewett along Shoshone River is:

	Feet
Cross-bedded, massive sandstone; thin beds of limestone	66
Greenish shale, sandy and glauconitic; belemnites and *Gryphaea*	300
Reddish shale with veinlets of gypsum	50
Granular to massive white gypsum	12
Fossiliferous limestone; interbedded shales	60
Fossiliferous sandstone	5
	493

Morrison and Cloverly formations (Lower and Upper [?] Cretaceous).—The separation of Morrison and Cloverly formations in Wyoming has customarily been made on the basis of a conspicuous conglomerate bed between two zones of varicolored shales. No such conglomerate occurs in the Shoshone River section; consequently Hewett includes all of the shales in the Morrison and leaves for the Cloverly only the rusty sandstones of the Greybull member. The writer, in mapping the Morrison and Cloverly in the Shoshone Canyon area, early concluded that no reliable criterion exists for differentiating the two formations. White sandstones in the Morrison are seen to lens out and be replaced by others at higher or lower horizons. A con-

[9] *Ibid.*, p. 268.

[10] D. F. Hewett, "The Shoshone River Section, Wyoming," *U.S. Geol. Surv. Bull. 541 (c)*, (1912).

glomerate character may appear in any one of these lenses and by no means marks a fixed horizon. The combined formations may be described in three groups:

	Feet
Buff sandstone, thin-bedded to massive (Greybull member)	110
Varicolored clays with many sandstone lenses	440
Green, sandy shales	180
	730

Thermopolis shale (*Upper Cretaceous*).—Following directly upon the Greybull sandstone are black shales of the Colorado group. That part which is known in the Bighorn Basin as Thermopolis shale consists of:

	Feet
Dark-gray, dense shale	50
Massive white sandstone	5
Black shale; thin beds of bentonite	390
Massive sandstone ("Muddy Sandstone")	20
Black shale	185
	650

Mowry shale (*Upper Cretaceous*).—The succeeding 375 feet of hard sandy shales in the Shoshone River section constitute the Mowry Formation.

Frontier sandstone (*Upper Cretaceous*).—The middle sandstone member of the Colorado group is known throughout northwestern Wyoming as the Frontier sandstone and is equivalent to the Torchlight sandstone of Montana. On Shoshone River it has a thickness of 495 feet.

Late Upper Cretaceous and early Tertiary formations.—No Cretaceous strata higher than the Frontier sandstone enter into the structural relations to be described, but they will be mentioned briefly to indicate the amount of post-Cretaceous erosion. The following formations are now recognized for the Cody region:

	Feet
Fort Union formation (Paleocene)	5,600
Lance formation (Paleocene)	1,800
Meeteetse formation (Upper Montana)	900
Mesaverde formation (Lower Montana)	1,200
Cody shale (Upper Colorado)	1,800
	11,300

Tertiary Channel deposits (Eocene ?).—Entirely distinct from the great series of strata which have been described are patches of arkose and conglomerate alined across the north end of Rattlesnake Mountain. These rest unconformably upon beveled Pennsylvanian and Permian strata. Granite, quartzite, and limestone boulders from the conglomerate are enveloped in the base of basalt flows at the head of Rattlesnake Valley. Thus the conglomerates are later than the first mountain uplift and earlier than the Miocene volcanics. From the arkoses, Dorf[11] has collected plants found in the typical Bighorn basin Wasatch beds. A reasonable interpretation is that these coarse materials mark a channel through which sediments were carried from mountain uplifts farther west and northwest to the basin of Wasatch deposition. Hence they represent a border phase of the Wasatch formation of Eocene age.

Terrace gravels and travertines in the vicinity of Cody have been well described by Fisher.[12]

Petrology of the pre-Cambrian.—Four distinct rock types are represented in the pre-Cambrian complex exposed in the Shoshone Canyon. In the probable order of age they are: hornblende-chlorite schist; quartz-diorite, granite-pegmatite, and diabase. The first three are intimately associated to form one gneissic complex, whereas the diabase cuts the latter with sharp contacts. Attenuated lenses and stringers of schist consist of hornblende, chlorite, quartz, and a little andesine. Absence of albite, biotite, or sericite suggests that these bodies represent xenolithic inclusions of an early paraschist.

Quartz diorite, constituting the main mass of the pre-Cambrian, consists essentially of quartz, oligoclase, and greenish biotite. Much of the diorite has been permeated by pegmatitic solutions which introduced microcline and additional quartz. Thick layers of granite pegmatite are emplaced along flat joints in the quartz diorite. Pure bodies of the pegmatite consist of pink microcline, quartz, and "nests" of mica.

Diabase dikes and sills intrude along joint planes of the gneissic complex. One immense dike in the bed of Shoshone River is over

[11] Erling Dorf, verbal communication.

[12] C. A. Fisher, "Geology and Water Resources of the Bighorn Basin," *U.S. Geol. Surv. Prof. Paper 53* (1907), pp. 34, 61, Pl. XVI.

100 feet thick and surrounds a large xenolithic body of granite. Petrographically it is an augite-diabase with a little quartz, resembling those common to all exposures of pre-Cambrian in Wyoming.

Tertiary igneous rocks (Miocene).—Flows and agglomerates of pyroxene-andesite extend across the head of Rattlesnake Valley onto Pat O'Hara Mountain. They represent a common type of the early basic volcanics of the Absaroka Range described by Iddings.[13] Two large dikes of andesite-porphyry parallel the west side of Rattlesnake Mountain and a third was discovered cutting the Cretaceous south of Ox Yoke Canyon. The fissure lines marked by these dikes show a close relation to the mountain deformation.

STRUCTURE

Faulting, flexing, and overthrusting of latest Cretaceous and early Tertiary age have produced the major structural features of the Shoshone Canyon area. In addition to these, there are indications of tectonic disturbances in the pre-Cambrian rocks alone, and of late Tertiary or Recent adjustments by faulting. These features will be interpreted in order of age rather than of importance.

Pre-Cambrian jointing.—The well-developed joints in the gneissic complex of Shoshone Canyon may be grouped with few exceptions into three systems: N. 5° E., δ 75° W.; N. 10° E.; and N. 90° E., δ 85° S. The flat-lying joints clearly were developed before or during the intrusion of the granite-pegmatites, and the steeper systems before the pre-Cambrian diabase. It is interesting to note that were the pre-Cambrian peneplained surface restored to a horizontal position by rotation about a N. 50° W. axis two of the joint systems would become vertical and the third nearly horizontal. The remaining joints in the pre-Cambrian are grouped adjacent and parallel to the fault plane west of Shoshone dam. This fault is of late Cretaceous age and it is reasonable to assume that the closely associated joints were contemporaneous.

Two significant deductions are possible. (1) During pre-Cambrian time orogenic forces produced systems of latitudinal, meridional, and horizontal joints followed by intrusions of pegmatite and dia-

[13] J. P. Iddings, "The Igneous Rocks of the Absaroka Range, etc.," *U.S. Geol. Surv. Mon. XXXII*, Pt. II (1899), pp. 275–96.

base. (2) When the next great orogenic cycle affected the region in late Cretaceous time the prevailing directions of faulting and jointing were at an angle of approximately 45° to those of the pre-Cambrian.

No evidence of orogenic activity in Paleozoic or early Mesozoic time has been discovered in western Wyoming. To be sure, numerous epeirogenic movements must be called into play to account for the several hiatuses and disconformities within the stratigraphic series.

LARAMIDE STRUCTURES

All the important faults and overthrusts of the Middle Rocky Mountain region were developed during one or more stages of post-Cretaceous or Laramide orogeny. To which stage the several events in each locality should be assigned is seldom certain and the exact geologic dating of any stage is usually not possible. Five stages are suggested by Hewett[14] for the west side of the Bighorn basin. Four recognized in the Shoshone Canyon area are: (1) pre-Wasatch faulting and "folding," (2) late Eocene overthrusting, (3) latest Eocene faulting, and (4) post-Miocene faulting.

Normal faults and flexures of Rattlesnake–Cedar Mountain.—The deep incision of Shoshone River reveals beyond any doubt the nature of the Rattlesnake–Cedar Mountain uplift. Pre-Cambrian granite is brought into contact with southwestwardly dipping beds of Ordovician dolomite by a steep normal fault. (Fig. 1, Sec. 3; Fig. 5). Erosion has nowhere reached pre-Cambrian rocks on the southwest side of the fault; hence, neither the exact displacement of the pre-Cambrian erosion surface nor the dip of this surface to the west can be observed. The displacement of the surface of the granite is probably greater than the 2,200 feet maximum displacement of the Bighorn dolomite; possible more than 3,000 feet.

Both northwest and southeast of the river the mountains rise rapidly, so that the fault trace passes entirely within the Paleozoic strata. The displacement is distributed among several branches of the fault on the north flank of Cedar Mountain, and is so largely taken up by yielding of the Lower Paleozoic sediments that Madison

[14] D. F. Hewett, "The Heart Mountain Overthrust, Wyoming," *Jour. Geol.*, Vol. XXVIII (1920), table facing p. 556.

limestone shows a nearly unbroken flexure below the mountain crest (Fig. 1, Sec. 4). In Cedar Mountain the uptilted mountain block terminates. Around the dip-slope of this mountain the strike of the Tensleep sandstone undergoes a change of 180°, of which 60° occurs abruptly at the southwest angle. The Embar and Chugwater formations likewise encircle Cedar Mountain as the uplift plunges to the southeast.

Half a mile northwest of the river, the great fault splits into two, which continue parallel for 7 miles along the face of Rattlesnake

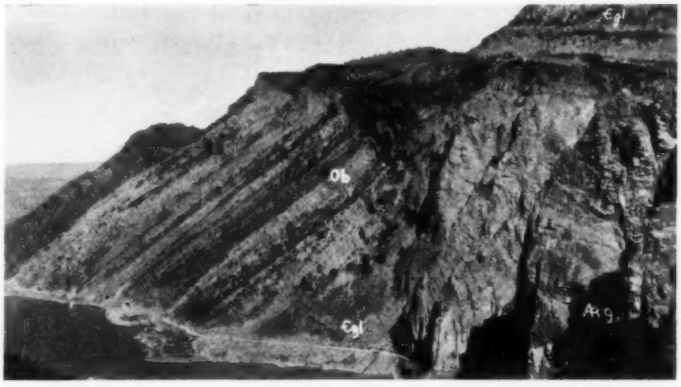

FIG. 5.—Displacement of Rattlesnake Mountain fault at Shoshone dam. Looking northwest from Cedar Mountain (cf. Fig. 1, cross section 3).

Mountain. Displacement by the inner or northeastern of the two faults becomes less than 1,000 feet, and that by the outer fault is greater but difficult to estimate, since it so closely parallels the bedding. The fault trace across gulches and spurs shows that the outer fault dips less steeply than the inner one; hence the two planes probably do not converge in depth. The tripartite structure is well exhibited in each transverse gulch (Fig. 1, Sec. 2; Fig. 6). The uptilted mountain block shows uniform northeast dips and rises in steplike cliffs which have weathered back from the fault plane. The median structural block, between the two fault planes, appears as a segment of a flat arch, the beds dipping gently toward each fault. Strata of the downthrow block dip 45°–90° to the west along the mountain

front but rapidly flatten out under Rattlesnake Creek Valley. Overturning of the beds, depicted in cross section 2, is restricted to less than 2 miles of the mountain front.

The aspect of the west side of Rattlesnake Mountain undergoes a distinct change beyond the deep gorge of Canyon Creek. The outer of the two parallel faults no longer reaches the surface, but in line of its projection is a right-angle flexure in the Madison limestone. In gulches erosion has cut to the upper beds of the Bighorn dolomite and on the spurs part of the fractured Madison limestone has been

FIG. 6.—Tripartite structure of Rattlesnake Mountain, 3 miles northwest of Shoshone dam. Looking southeast (cf. Fig. 1, cross section 2).

removed along the axis of the flexure. The resulting topographic expression is a series of palisades of vertical Madison limestone and a cliff marking the edge of the same beds in a horizontal position east of the flexure (Fig. 1, Sec. 1; Fig. 7). A second less pronounced cliff farther east indicates that the inner fault continues northwestward with diminishing displacement at the surface. As the land rises toward the head of Rattlesnake Creek, the depth to which the structure is exposed decreases. The steep west limb and flat east limb of the Rattlesnake Mountain flexure, however, can be traced practically to the edge of the volcanic cover.

The northeast slope of Rattlesnake Mountain is not precisely a uniformly dipping structural plane. Flat dips along the crest in-

crease eastward to a maximum of 20°, then flatten and steepen again to form a structural bench at an elevation of about 6,500 feet. Farther northeast the dip decreases to a gentle syncline. A few small, transverse elements were observed in the mountain structure. A small outcrop of pre-Cambrian attracted attention to a fault of about 150 feet displacement, northwest of Shoshone Canyon. Evidence of a fault of the same sort in the bed of Canyon Creek is not conclusive. Near the head of Trail Creek a small syncline, transverse to the general strike, causes the outcrops to be deflected.

FIG. 7.—View northwest to head of Rattlesnake Creek. Overthrust Madison limestone of Chalk Mountain (left). Relatively displaced remnants of overthrust (center). Vertical and horizontal Madison of Rattlesnake Mountain (right) (cf. Fig. 1, cross section 1).

A flat uplift in the Mesozoic strata, southeast of Cedar Mountain, would appear to be related to the Rattlesnake Mountain structure. The two structural axes, however, differ in strike by nearly 45°. Although the east limbs of the structures are continuous from one to the other, the hogback ridges southwest of Cedar Mountain are offset 3 miles to the east by two sharp bends and a strike fault. The strata rise in steep hogback ridges on either side of the structure, connected by a very flat arch, slightly depressed at the crest.

Pat O'Hara Mountain uplift.—Pat O'Hara Mountain is a structural unit distinct from the Rattlesnake–Cedar Mountain uplift. Its east-west axis diverges at 45° from the north end of Rattlesnake Mountain. It is a similar uptilted block with a steep fault on its south flank and moderate dips northward from the crest. The gentle dip slope of Rattlesnake Mountain continues northwestward to within less than a thousand feet of the east-west strikes of Pat O'Hara Mountain. Embar and lower Chugwater beds abut on the Pat

O'Hara fault, whereas the upper Chugwater, Sundance, and Cloverly turn sharply through an angle of 120° to continue with vertical dips parallel to the Pat O'Hara axis. Thus there is produced between the two mountain uplifts an asymmetric syncline with the axis pitching S. 80° E.

On Pat O'Hara Mountain the fault which reaches the surface is well below the crest of the uplift. Its stratigraphic throw is small, compared to the displacement of over 1,000 feet, since it parallels the strike and dip of the beds. The fault passes to the east into a mass of landslide débris where it probably terminates against a cross fault. Toward the west the Pat O'Hara fault is last evident where Madison limestone is brought in contact with Chugwater sandstones. Beyond that point the intersection of the Rattlesnake and Pat O'Hara Mountain structures is obscured by volcanic rocks and their débris, as well as by soil cover on the flat upland. Scattered outcrops show strikes and dips discordant with either of the mountain structures, or show intense crushing which has obliterated the bedding. The most conspicuous of these isolated outcrops shows a definite east-west strike and dip of 45° to the north.

Heart Mountain overthrust.—Along the west edge of Rattlesnake Creek Valley are cliffs of Paleozoic limestone which are known to be part of the Heart Mountain overthrust sheet. That part of the thrust plane shown in Figure 1 is everywhere covered by talus, but its approximate trace indicates a nearly horizontal dip. Generally the limestone of the overthrust block is fractured and its dip, where discernible, is irregular in direction but nowhere steep. Frontier and Mowry beds underlie most of the overthrust west of Rattlesnake Creek, but a small segment near the head of the creek is definitely in contact with Cloverly and Sundance. Large blocks of the limestone appear to have moved down the slope by creep, especially at the township line 52N.–53N.

All known or probable remnants of the overthrust sheet are shown in Figure 2, and in Figure 7 the relation of overthrust Madison limestone to the same beds involved in the Rattlesnake Mountain structure is clearly shown. In the original map of the overthrust Dake[15] shows secondary thrust faults beneath the principal plane. None of

[15] *Op. cit.*, p. 48.

these is within the bounds of the present map, but structures believed to be similar are shown southwest of Cedar Mountain and in the upper Rattlesnake Valley.

INTERPRETATION OF STRUCTURE

Block uplifts.—The essential feature of the Rattlesnake–Cedar Mountain uplift is apparent from a geologic map and cross sections. Uplift and tilting of one large block of basement complex are fundamental, and to these flexing and minor faulting are sequential. To account for the abrupt termination of the uplift in Cedar Mountain it would be postulated that the uplifted basement block is bounded by an east-west fault beneath the crest of that mountain. Such a fault would be rotatory about a horizontal axis and would not correspond to the strike fault which outcrops some distance farther south. What appears to be a continuation of the same uplift to the southeast is probably a second structure of similar type but of less magnitude. Sharp bends in the outcrops south of Cedar Mountain represent the adaptation of strata of varying resistance to the two adjacent displaced blocks beneath.

Fluctuations in the dip and local overturning of the beds along the west flank of Rattlesnake Mountain are comparable to similar conditions on the flanks of the Bighorn Mountain uplift. The development of such variations along the Colorado Front Range is sketched by Finlay.[16] It is doubtful if such changes along the front of an uplifted mountain range indicate any fundamental difference in structural conditions.

Pat O'Hara Mountain is nowhere sufficiently incised to reveal directly its deeper structure. The presence of a normal fault parallel to the axis of uplift, the thinning of the upturned strata, and the abrupt termination of the uplift on the east indicate that its structure is of the same type as Rattlesnake–Cedar Mountain. The number and attitude of displaced basement blocks underlying Pat O'Hara Mountain must be somewhat different to have produced a bowed rather than sharply flexed form of the Madison limestone over the crest (Fig. 1, Sec. 1).

[16] George I. Finlay, "Colorado Springs Folio," *U.S. Geol. Surv. Folio 203* (1916), Fig. 7.

Finally, what conclusions can be drawn as to the way in which the two uplifted structures intersect? Two indications that the structure of Pat O'Hara Mountain was developed, at least in part, after that of Rattlesnake Mountain are: (1) termination of the northeast dip slope against the Pat O'Hara fault, and (2) the presence of eastward dipping beds of Madison limestone west of Pat O'Hara Peak, which could be interpreted as an uplifted portion of those of Rattlesnake Mountain.

The normal fault structures of Rattlesnake and Pat O'Hara mountains are in many respects similar to fault blocks of the "plateau" or "basin-range" type. Gilbert[17] in his original description of the type shows in a series of cross sections of the Sevier fault, the lateral and vertical gradations of a normal fault into a sharp flexure and its branching into a group of sub-parallel faults. Many examples of the gradation of faults into flexures have been recognized and a number of geologists have discussed the relation of deep-seated faults to overlying structures. In particular, the monoclinal flexures and flat-topped uplifts in the Cretaceous of central Montana have been interpreted by Thom[18] as the result of faulted basement blocks. There the faulted blocks are merely interpreted from drill records and geophysical data, but in Shoshone Canyon the deeper portions of just such a structure are remarkably revealed by a relief of over 4,000 feet. Every stage from the upfaulted block of pre-Cambrian granite to the sharply flexed Carboniferous strata is exposed to observation. The final expression of the uplift in the higher Cretaceous formations is not shown in Rattlesnake Mountain; but, with this assurance of the correctness of Thom's interpretation, the "anticlines" of central Montana and of the Bighorn Basin are believed to be precisely this expression.

Flexures of the type so well exemplified in Rattlesnake Mountain have been variously named "incompetent folds," "plains-type folds," or "*unechte Falten*." They possess certain characteristics which

[17] G. K. Gilbert, "Geology of a Portion of Nevada, Utah, California and Arizona," *U.S. Geol. Surv. west of 100th Mer.*, Vol. III (1875), p. 50, Fig. 25.

[18] W. T. Thom, Jr., "The Relation of Deep-seated Faults to Surface Structural Features in Central Montana," *Bull. Amer. Asso. Petr. Geol.*, Vol. VII (1923), pp. 1–13.

distinguish them from true folds of the Appalachian type, and indicate their similarity to monoclinal flexures. (1) Anticlinal flexures never occur as a wave-like succession of anticlines and synclines but as individual uplifts above the regional level of the strata. (2) Anticlinal flexures are usually distinctly asymmetric; or, if symmetrical with respect to the axial plane, show a flat top bordered by sharp flexures. (3) Normal faults parallel to the axes of flexures are abundant, whereas in competently folded regions normal faults are less conspicuous and more often transverse. (4) Thickening of strata does not occur, and thinning takes place near the axis of flexures rather than on the limbs. (5) The axial planes of asymmetric, incompetent folds tend to be vertical or dip toward, rather than away from, the steep limb.

Certain striking features of a large number of anticlines in the Bighorn Basin were noted by Hewett and Lupton.[19] With very few exceptions the anticlines occur in a zone bordering the Basin; each axis strikes roughly northwest; and each uplift is noticeably asymmetric in cross section. Furthermore, the asymmetry is consistently such that the flatter limb of the flexure is toward the center of the basin, whether that be northeast or southwest. Rattlesnake Mountain meets every condition for the largest of the border-zone anticlines. Many, if not all, of the other anticlines show features and associations characteristic of incompetent rather than competent folds. With definite knowledge of Rattlesnake Mountain, the writer would suggest that each anticline of the border-zone marks a displacement in the basement complex by uplift and tilting. Two reasons for the absence of anticlines in the center of the Bighorn Basin are (1) that uplifts were largely pre-Wasatch and are still covered by unconformable sediments, and (2) that small displacements in the basement complex would be completely absorbed by the greater thickness of soft or unconsolidated strata overlying.

Whatever may be the relationship of Rattlesnake Mountain to the Beartooth Range, it is certainly not one of simple continuity. A detailed study of the intervening Sunlight basin area may ultimately

[19] D. F. Hewett and C. T. Lupton, "Anticlines in the Southern Part of the Bighorn Basin, Wyoming," *U.S. Geol. Surv. Bull. 656* (1917), Pl. 1, p. 33.

indicate some connection between the two uplifts, and show that Rattlesnake and Pat O'Hara mountains are parts of the enclosing ranges rather than local uplifts within the Bighorn Basin.

Heart Mountain overthrust.—No new or conclusive evidence as to the direction, magnitude, or method of emplacement of the Heart Mountain overthrust was gained. Two bodies of limestone, however, one south and one west of Pat O'Hara Peak (Fig. 1; Fig. 7), when interpreted as remnants of the overthrust, give a definite measure of later displacement of the thrust plane. These two remnants are not shown on earlier maps of the overthrust. Their discordant strikes and dips, and excessive fracturing, however, suggest that they should be included. The limestone body south of Pat O'Hara Peak is preserved by its protected position on the down-throw side of the Pat O'Hara fault, whereas all traces of the overthrust have been removed from the higher ridge of Rattlesnake Mountain.

In order to interpret the geologic history of the region, and the relations to Rocky Mountain orogeny, one must accept a hypothesis for the development of the Heart Mountain overthrust. Since Hewett[20] suggested that the Heart Mountain overthrust rode forward over an even surface of erosion and aggradation, other conclusive examples of "erosion thrusts" have been described by Longwell, Hume,[21] and others. Moreover, it is believed that the Heart Mountain thrust belongs to the same phase of orogeny as the great zone of overthrusts which extends from southeastern Nevada to northern British Columbia. Even though Heart Mountain lies well to the east of this northerly trending zone, it is not thought necessary to invoke any singular cause of overthrusting such as that suggested by Bucher.[22]

Post-overthrust faulting.—Evidence of renewed uplift of Rattlesnake and Pat O'Hara mountains after overthrusting is not entirely conclusive. It offers, however, the simplest explanation of the height to which Rattlesnake Mountain rises above the overthrust on either

[20] *Jour. Geol.*, Vol. XXVIII (1920), p. 554.

[21] C. R. Longwell, "Structural Studies in Southern Nevada and Western Arizona," *Geol. Soc. Amer. Bull.* 37 (1926), pp. 570–72; G. S. Hume, "Overthrusting in the Foothills of Alberta," *Econ. Geol.*, Vol. XXVI (1931), pp. 263–68.

[22] W. H. Bucher, "Volcanic Explosions and Overthrusts," *Trans. Amer. Geophys. Union for 1933*, pp. 238–42.

side. It accounts for the different levels of overthrust remnants north, west, and south of Pat O'Hara Mountain. Wasatch conglomerates appear to be down-dropped at the head of Rattlesnake Valley and the arkoses are cut by small transverse faults. Normal faults displacing the overthrust planes are conspicuous features of areas described by Mansfield and Hewett,[23] who interpret them as mid-Tertiary or "basin-range" in age. This age for the Rattlesnake Mountain displacements seems very plausible, since it would group them with the faulting which must necessarily have accompanied the intense Miocene vulcanism in the Absaroka Range. On the other hand, there is a suggestion that faulting has continued until recent time.

At the headwaters of Rattlesnake Creek, within an area of half a square mile (blank in Fig. 1), the soil is heaved and crumpled. Vertical displacements along approximately straight lines and the absence of any landslide scars suggest a direct tectonic disturbance. Moreover, in June, 1925, open fissures developed in the lower part of Rattlesnake Valley and an abandoned well near Cody depot suddenly gushed forth. What relation exists between these phenomena and displacements on the Rattlesnake and Pat O'Hara faults was not ascertained.

Types of Rocky Mountain structure.—Many geologists have recognized distinct types of structure in various parts of the Rocky Mountains. Chamberlin[24] refers to them as "thick-shell" and "thin-shell" types of mountains and suggests a distinction between the northern and the southern Rockies. Penck[25] elaborates the discussion further and rather overemphasizes the demarcation of the Rocky Mountains of Montana and Canada from those of Wyoming and Colorado. Thrust structures, to be sure, predominate over simple block uplifts in the Canadian and Montana Rockies, but through western Wyo-

[23] G. R. Mansfield, "Geography, Geology and Mineral Resources of Southeastern Idaho," *U.S. Geol. Surv. Prof. Paper 152* (1927); D. F. Hewett, "Geology and Ore Deposits of the Goodsprings Quadrangle, Nevada," *U.S. Geol. Surv. Prof. Paper 162* (1931).

[24] R. T. Chamberlin, "The Building of the Colorado Rockies," *Jour. Geol.*, Vol. XXVII (1919), p. 251.

[25] A. Penck, "Geomorphologische Probleme in ferner Westen Nordamerika," *Sitzber. Preus. Akad. Wissen.* (1929), *Math.-Phys. Kl.*, pp. 207–9.

ming, eastern Idaho, and Nevada is a zone of overthrusts of scarcely less magnitude. Kober[26] takes cognizance of the continuity of the zone of overthrusts in developing his schematic picture of orogeny of North America. Finally, Billingsley[27] makes a sharp distinction between the "overthrust arc" and the "folded zone" of the Rocky Mountains.

The writer also believes it more objective to divide the Rocky Mountain structures into longitudinal zones rather than a northern and a southern type; but he would not go so far as to deduce the symmetrical orogenic plan of Kober, nor would he make such broad interpretations as does Billingsley.

It has frequently been noted that there is a progressive thickening of the Paleozoic strata from east to west in any section across the Cordilleras. The important fact from a structural point of view is that within the zone of large overthrusts the pre-Cretaceous strata are, with few exceptions, over 7,500 feet thick. In the eastern region of uplifted ranges the corresponding column is incomplete and rarely more than 2,000 feet thick. The Paleozoic series in Montana is less thick, but the Algonkian Belt series plays an equivalent rôle in the overthrusting. A relation between thick series of geosynclinal sediments and intensive compressional deformation has been recognized by nearly all students of orogenesis. The writer merely wishes to emphasize the widespread application of this association in the Cordilleran region. What were the fundamental determining factors for the two types of Rocky Mountain structure is entirely conjectural.

This somewhat extended discussion of the two major zones of Rocky Mountain orogeny is justified by the fact that within a small area in northwest Wyoming the two overlap. The greatest extent of overlap would be from the overthrust on Heart Mountain to the uplift of the Teton Range. Evidence of the distinct ages of the two types of deformation has been found in the Shoshone Canyon region. Hewett[28] concluded, from the age of the latest overridden sediments and the oldest overlying volcanics, that overthrusting was during

[26] L. Kober, *Der Bau der Erde*, 2d ed. (1928), pp. 316–21.

[27] Paul Billingsley, "Geologic History and Geologic Structure of the Salt Lake Region," *XVI Int. Geol. Cong. Guidebook 17* (1933), pp. 19–24.

[28] *Op. cit.*, p. 555.

late Eocene. It has been shown that the uplift and beveling of the Rattlesnake Mountain structure was not merely pre-overthrust but also pre-Wasatch. Materials in the Wasatch, unmistakably derived from the pre-Cambrian complex, indicate that great uplifts to the west and north had risen and been deeply eroded before the Eocene. One of these uplifts was the Beartooth Range against which the later Heart Mountain overthrust probably impinged. The age of the Bighorn Mountain or Owl Creek Mountain uplifts is not determined. The Laramie and Colorado Front Ranges are usually interpreted as late Cretaceous, and one might believe that all the larger pre-Cambrian blocks had started to rise before Wasatch time.

There may be represented in western Wyoming yet a third phase of Cordilleran orogeny which is typified by basin range structure, i.e., mid-Tertiary normal faulting. Whereas in Nevada and Utah these faults formed fresh breaks across both folded and overthrust structures, in the Shoshone Canyon area they resulted in renewed movement along the pre-Eocene fault planes of the underlying basement.

GEOLOGIC HISTORY

A few traces of the earliest recorded stage of the geologic history of the region are preserved in the lenses of hornblende-schist of the pre-Cambrian complex. Early pre-Cambrian sediments, which are more extensively preserved in neighboring areas, have been altered to schist and para-gneiss during a period of orogeny. This was followed by intrusion of large bodies of granitic rock together with more basic intrusives. Joints developed during this intrusion were filled without interval by pegmatites. At a distinctly later period in the pre-Cambrian, basaltic dikes intruded all the older complex throughout the region. A long cycle of erosion, in late pre-Cambrian and early Cambrian time, produced a nearly plane surface across all of the early rocks. This surface is everywhere irregularly uplifted or tilted but is rarely seen to be sharply folded or warped by later orogeny.

The sea advanced over this erosion surface in the middle Cambrian, receded in the early Ordovician, and readvanced to greater depth in late Ordovician time. Before the end of the Silurian the region was raised above sea level so that channels were eroded and later filled by Devonian sediments. The slight sedimentation during the

Devonian was followed by uniform marine conditions in Mississippian time. In the Pennsylvanian, frequently changing conditions of sedimentation produced a series of limestones, shales, cherts, and cross-bedded sandstones. Continental and peculiar shallow-water conditions prevailed through the Permian and Triassic. Following a brief emergence in the early Jurassic, normal shallow marine conditions continued through late Jurassic, early Cretaceous, and a large part of late Cretaceous time.

Although evidence exists that regions to the west and north were subject to orogenic disturbances as early as the Jurassic, the first uplift of the Rattlesnake Mountain structure is believed to have taken place very late in the Cretaceous. Eocene erosion truncated the late Cretaceous uplifts, cutting down to the Carboniferous limestone on Rattlesnake Mountain and to the pre-Cambrian granite in the mountains which had risen farther to the northwest. Subsidence of the Bighorn basin continued through the Eocene and large amounts of material were transferred from the elevated regions of Paleozoic and pre-Cambrian exposures to the Basin, where they were deposited as Wasatch arkoses and clays on top of the thick Cretaceous sedimentary series. Rattlesnake Mountain lay at the border of the basin of deposition where a small thickness of coarse material was deposited along stream channels.

Upon the Wasatch surface, made even by erosion and aggradation, the great mass of Paleozoic limestones of the Heart Mountain overthrust plate moved forward. New lines of drainage established themselves upon the surface of the overthrust plate. During the Oligocene and early Miocene these rivers removed a large part of the fractured limestone of the upper block and had begun to incise the underlying surface, developed in the early Eocene. The major streams maintained the courses which they developed on top of the thrust block and cut indiscriminately into the underlying structures. Thus, Shoshone River in the Shoshone Canyon is clearly superimposed upon the Cedar–Rattlesnake Mountain uplift. Another definite example of a superimposed stream within the Heart Mountain overthrust zone is Ox Yoke Canyon of Pat O'Hara Creek.

Following the overthrusting and possibly continuing to Recent times there has been further differential elevation of the Rattlesnake

Mountain and Pat O'Hara Mountain blocks by renewed movement along the pre-Eocene lines of faulting. On the basis of the present positions of remnants of overthrust limestone and of Wasatch conglomerate, the renewed uplift of Rattlesnake Mountain is estimated to be 1,200–2,000 feet. The west end of Pat O'Hara Mountain and the South end of Rattlesnake Mountain were the most elevated. Consequently there is a differential uplift of 800–1,600 feet of Pat O'Hara Mountain above the north end of Rattlesnake Mountain. During this uplifting, Shoshone River maintained its course as an antecedent stream.

By late Miocene most of the overthrust plate had been removed when great quantities of "early basic breccia" were ejected to form a new mountain range of the Absarokas. Volcanics are found at an elevation of 7,500 feet in Rattlesnake Valley and at 10,100 feet on Pat O'Hara Peak. Probably they once covered much of Rattlesnake and Pat O'Hara mountains and a large part of the soft Cretaceous and Eocene beds south and west of Cedar Mountain, but there is some doubt whether the site of Shoshone Canyon was under lava cover.

Since the Pliocene, erosion has dissected the lava plateau of the Absarokas as rapidly as the stream gradient has been renewed by the cutting of Shoshone Canyon. Where soft Cretaceous and Tertiary beds were unprotected or capped by only a thin layer of lavas or overthrust limestones, broad valleys were opened out. Especially to the south of Cedar Mountain a broad plain has been eroded and two streams flowing into the Shoshone River above and below the Canyon, respectively, are separated by a very low divide. Rattlesnake Creek and Trail Creek developed as subsequent streams along the strike of the softer strata and help to emphasize the topographic expression of the Rattlesnake Mountain uplift.

Rock benches and terraces indicate the later stages of development of Shoshone Canyon, but neither moraines nor glacial outwash are found in the immediate vicinity. The oldest feature is a bench on either side of the canyon, 1,000 feet above the present stream. It is not developed directly upon the surface of the resistant pre-Cambrian granite, but 300–500 feet stratigraphically above it, indicating a broad valley developed at a determined elevation without regard

for rock hardness. Correspondence in elevation with the broad, outer valley of Clark Fork is very close. That level is considered by Alden[29] to be the mountainward projection of the Flaxville plain of Pliocene age.

Upon the Flaxville surface, before its dissection, there are, in a few Montana localities, remnants of early Pleistocene moraines. During the Pleistocene but before Wisconsin time, there were two more stages of river entrenchment and valley broadening. The first of these is represented east and south of Shoshone Canyon by broad gravel terraces 200–300 feet above the present streams. This level skirts the base of Carter Mountain and extends down both sides of Shoshone River above the town and station of Cody. Gravel-capped hills west of Sulphur Creek are believed to belong to this stage, since the level at which the smaller creeks are now flowing corresponds to the next younger terrace of the river. The innermost river terraces, 100–150 feet above stream level, are shown by Alden to underlie the Wisconsin moraines. Irma Flat, southeast of Shoshone Reservoir, and the terrace on which the town of Cody is located represent this stage. In Recent times the Shoshone has entrenched itself in the underlying bed rock east of the canyon and has continued to deepen its gorge through the granite core of Rattlesnake Mountain. Entrenchment of the late Pleistocene terrace is proceeding headward along all of the tributary streams with the exception of Rattlesnake and Diamond creeks where normal development has been halted by the artificial water level of Shoshone Reservoir.

[29] W. C. Alden, "Physiography and Glacial Geology of Eastern Montana and Adjacent Areas," *U.S. Geol. Surv. Prof. Paper 174* (1932), p. 27.

ROUNDED LAKES AND LAGOONS OF THE COASTAL PLAINS OF MASSACHUSETTS

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ABSTRACT

There are many lakes and lagoons in the sandy areas along the southeast coast of Massachusetts, which show a tendency to develop rounded shores of more or less oval outlines with their long axes in the direction of the greatest wind velocities. These forms are here regarded as results of normal shore processes, primarily of wave erosion, but they can fully develop only under especially favorable circumstances. Possible relationship to the supposed meteorite scars of South Carolina is discussed.

INTRODUCTION

The following study is based on the airplane survey of Massachusetts, completed in 1933 by the Institute of Geographical Exploration of Harvard University, in connection with a course in aerial photograpy. The entire state has been photographed by Lieutenant J. F. Phillips and Mr. Weld Arnold, with a Fairchild 5-lens camera from an altitude of 15,000 feet, resulting in 4,000 photographs at an approximate scale of 1:30,000.

The study of the photographs from a geological point of view was very profitable; particularly useful were the pictures for the study of shore processes, as water bodies show up well in airplane photographs and the bottom can be seen down to several fathoms.

LAKES AND LAGOONS

One feature stands out particularly conspicuously in these photographs: the rounded outline or often cuspate form of most of the lakes and lagoons in the sandy coastal regions. The study of several hundred lakes and lagoons from photographs and in the field led to the following observations:

1. The originally irregular shores of lakes and lagoons tend to develop rounded forms and more or less circular outlines. They can be found in all stages of development.

2. The shores of more elongated lakes and lagoons tend to develop a scalloped outline consisting of round, slightly oval bays, separated by cuspate points. This process can be observed in non-tidal as well as in tidal waters.

3. The radii of the rounded bays are in rough proportion to the width of the water body, the ratio varying from one to two.

4. The V-shaped cuspate points sometimes enclose a small lagoon or marsh, but in the majority of the cases this is absent.

5. These developments are far more advanced in sandy ground than in clay, gravel, or mixed material.

6. Lakes and lagoons under 1,000 feet of width do not show well the listed characteristics. Few lagoons involved in this study exceed 2 miles in width.

The following illustrations may bear out some of the preceding features.

LAKES WEST OF WELLFLEET, CAPE COD

These lakes were probably originally ice-block holes in glacial outwash, and are nearly at sea level. Gull Pond, Great Pond, and some of the other lakes are remarkably round.

Gull Pond, looking south toward its steep, 60-foot bluffs, has a distinct crater-lake appearance. When we look, however, in the northerly direction, this impression immediately disappears, because the narrow bar which separates it from the lake to the north is a shallow sand bar, only 3 or 4 feet above the water level. As the same feature is repeated almost exactly between the two northern lakes, little doubt exists that Gull Pond, and the lakes north of it, once formed one great irregular lake which became separated by sand bars. It should be noted that in the curvature of the sand bar north of Gull Pond the arc of the larger lake is dominant. The white spot in the middle of the bar is an artificial canal. The bluffs of the lake have all characteristics of wave-cut cliffs, and are supposed to have supplied the sand for the building of the bars. It is remarkable that the waves of such small lakes can produce cliffs 60 feet high, but it must be considered that the cliffs are of coarse sand and a 10-foot terrace gives evidence that the lakes may once have had a higher level; it is also possible that the initial shoreline may have been quite steep.

In Great Pond four former arms of the lake became separated by sand bars which are so low that at the time of the author's visit they were partially submerged owing to flooded conditions. Great Pond has a kidney form which is characteristic of many lakes of the region, not elongated enough to form a series of lakes as the Gull Pond group, and not compact enough to develop into single oval lakes.

FIG. 1.—Rounded Lakes, Massachusetts

Whether this lake in its future cycle will be separated into two lakes by a central bar or a single oval lake depends on the amount of sand available for the building of a cuspate bar.

COSCATA CUSP AND COATUE BEACH, NANTUCKET ISLAND

1. Great Point, 2. Truncated Beach Ridges, 3. Coscata Pond, 4. Haulover Neck, 5. Coatue Beach, 6. Bass Point, 7. Polpis. (Fig. 2.)

A composite photograph or mosaic of Coscata Cusp of Nantucket Island shows a remarkable development of a tidal lagoon with a series of rounded bays and cuspate points. The two shores of the lagoon are very different. The southern shore is of mixed glacial material, with cliffs rising as high as 50 feet, whereas the other side consists of a low sand bar—a recurved spit in morphologic terminology. Because of this difference the development of cuspate outlines is far more advanced on the northern side. The continuation of the cuspate points can be traced under sea for a great distance. They are much better visible directly from the airplane as the photographs suggest. Farther north the perfection of curvature and the length of the submarine sand bar show stages of progressive development. In Bass Point the submarine sand bar reaches across the entire lagoon and a slight indication is seen on the other shore of the development of a counter point (cuspate point, developed opposite a projection on the other shore).

Some controversy has occurred as to the origin of the remarkable cuspate outline of Coatue Beach. Shaler has suggested tidal whirlpools.[1] Gulliver attributed it to eddies caused by the in- and outflowing tides.[2] Johnson regards it as the result of normal development of a recurved spit modified by the rounding effect of small waves of the lagoon.[3] Examination of aerial photographs shows that tidal currents could have played only a minor rôle in the development of the cuspate points, because the submarine bars, as they are, would rather obstruct tidal eddies than be formed by them. As

[1] N. S. Shaler, "Geology of the Island of Nantucket," *U.S. Geol. Surv. Bull. 53* (1889), p. 13.

[2] F. P. Gulliver, "Nantucket Shorelines," *Geol. Soc. Amer. Bull. 15* (1904), p. 521.

[3] D. Johnson, *The New England-Acadian Shoreline* (New York: Wiley, 1925), pp. 441–46.

FIG. 2.—Coscata

cuspate points are just as common in non-tidal as in tidal waters—so long as they are in low sandy ground—little support is found for Gulliver's theory. Johnson's view is well exemplified by the two southernmost and most recently developed points where the recurved ends of the progressive spit are distinctly visible. No such recurving ridges can be seen in the more northern points, but we may assume here that the shore has been cut back beyond the recurved edges, because the beach ridges at the northern end of the lagoon indicate that the spit may have been originally quite wide.

Note the cuspate outline of both basins of Polpis Harbor.

COASTAL PLAIN NEAR FALMOUTH, MASSACHUSETTS

Figure 3 represents a partially submerged outwash plain in front of a moraine, and it is shown primarily to illustrate the relationship between the size of rounded bays and the width of the water body. It seems that with less than 1,000 feet of fetch, the effect of the waves is too small to form cuspate points.

CYCLE OF WAVE EROSION IN LAKES AND LAGOONS

The origin of the rounded lakes and cuspate points can be well explained by normal processes of wave and current action as they are known to work on the seashore. These forms are not so common in lakes and lagoons, because the waves are small and the currents but feeble, and cuspate points develop only where enough sand is available. Lakes are not common in sand; they are possible only if the water of the lake cannot escape underground on account of high ground-water level. In clay, gravel, and rock the process of wave erosion is so slow that the lake is likely to die a natural death by filling and draining before the wave cycle is completed.

If the initial lake is fairly round and deep, the cycle of erosion will be similar to that of the cycle of the shore line of submergence. The waves will cut cliffs into the opposing shore, reducing the outstanding projections. The washed-out sand will be deposited partly in front of the cliff, building up a beach—corresponding to the shore profile of equilibrium—and in part it will be carried sideways by the process of beach drifting, thus building sand bars or spits. The sand bars may eventually reach across the shallow embayments of the

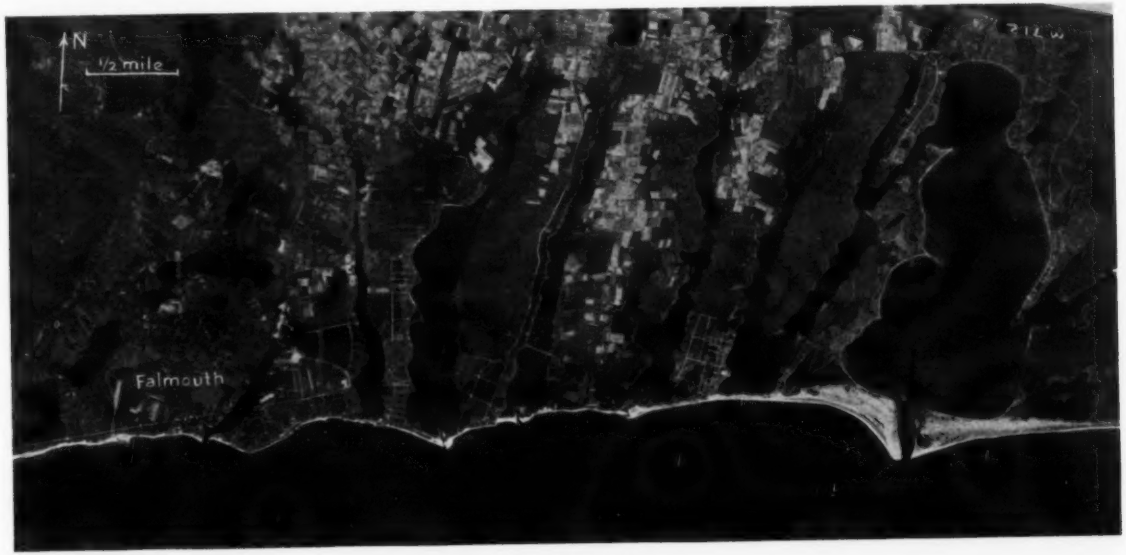

FIG. 3.—Falmouth

lake, separating them from the main lake. When a continuous shore of cliffs and sand bars is developed all around the lake, a stage is seen which is comparable to the stage of submaturity on a marine shoreline of submergence. This stage is represented by Great Pond. If the wind blows from all directions, the lake will become nearly round, though longest in the direction in which the original lake was elongated. If, however, there is a prevailing direction of storm winds, the leeward shore will be driven back faster than the windward shore and the resulting lake will be oval, oriented with its blunt end in the direction toward which the strongest winds blow. Ordinary waves on a small lake are very ineffective in producing changes after a profile of equilibrium is established.

In lakes originally irregular but long and narrow in outline the process is somewhat different. Waves, large enough for effective cutting of the shore, will be generated only in places where the lake is wider than about 1,000 feet. The waves have the longest fetch in the largest embayments and thus the embayment will be cut back faster then the projecting headlands. This is quite contrary to the development on open shores, where there is but little difference in the effective fetch of the waves and as a result the headlands are cut back first. In long and narrow lakes and lagoons the larger embayments will be widened and rounded by the waves cutting back the smaller projections within the larger embayment and closing shallow arms. The cuspate point which will remain between two rounded bays will be the natural place for hydraulic currents which carry back the waters piled up by strong winds. These currents carry much sand raked up by the agitated waters and by depositing this sand the cuspate point will grow seaward rather than recede.

It is beyond the scope of this paper to go into details of the formation of cuspate forelands; it is enough to refer to D. W. Johnson's book, *Shore Processes and Shore Line Development*,[4] in which the process which develops cuspate points and the counter points on the other shore is critically discussed. Though Johnson attributes the building of counter points in most cases to the shadowing effect of the projections, causing shorter fetch of waves, airplane pictures in-

[4] Pp. 319–39.

dicate that occasionally sand may be carried at the bottom straight across the lake, thus initiating the building of counter points. A similar process was suggested by Johnson in the development of baymouth bars. It should be noted that counter points are much more common in lakes than in lagoons where tidal currents are likely to hinder their developments.

If the lake is originally very shallow the development will be similar to that of a shoreline of emergence. In the central deeper part of the lake, larger waves will be generated which, breaking against the shallow borders, will pile up off-shore bars, the highest bars of which are near the shore facing the strongest winds. A continuous ring of off-shore bars around the deeper part of the lake will develop, enclosing a more or less circular lake of some depth, and cutting off the shallow arms, which, separated from the central lake, are likely to be filled with bog. In further development, the bar facing the strongest winds will be driven back the fastest, and an oval lake will develop. No lake was found in Massachusetts belonging exclusively to this type.

As for possible light which these developments may throw on the supposed meteorite scars of South Carolina, it should be stated that the rounded lakes of Massachusetts are far from being so perfect in shape, but these studies point to a possible method by which such craters may have been formed, somewhat as suggested by C. Wythe Cooke,[5] although not necessarily involving all elements of his theory. The chief difficulty with this explanation is that the direction of strongest winds in South Carolina seems to be transverse to the elongation of the "bays." The possibility must be considered, however, that the wind direction in the past may have been different. Considerable diversity exists in the winds of Massachusetts; on Cape Cod the strongest winds come from the northwest, whereas on Nantucket Island, only 30 miles away, they come from the northeast.

The following criteria may be helpful in deciding whether the Carolina Bays are meteorite craters or the results of normal shore processes:

[5] "Discussion of the Supposed Meteorite Scars in South Carolina," *Jour. Geol.*, Vol. XLII (1934), pp. 88–104.

In Case of Normal Wave-cycle Development	In Case of Meteorite Scars
1. Round lakes will be best developed in sandy ground.	1*a*. Craters are possible in any unconsolidated material.
2. Rim is built of washed sand.	2*a*. Rim is of the same kind of material as the crater.
3. Rim (sand bar) is developed on low portions only. Absent where the shore is cut into higher land.	3*a*. Rim is present even where the crater is cut into higher land.
4. Many lakes will exist in echelon arrangement with a narrow sand bar between.	4*a*. No preference to this arrangement.
5. Transitional forms representing earlier stage of development are probable.	5*a*. No transitional forms.
6. Lakes are most common in bottom lands or flat uplands; not possible on well-drained slopes	6*a*. Distribution of craters or their remnants has no relation to the topography.

Use of these criteria in a further study of the South Carolina bays may throw some light on their origin.

SIR T. W. EDGEWORTH DAVID[1]

GRIFFITH TAYLOR

The "Grand Old Man" of Australian Science died on August 28, 1934, after half a century of work on the geology of the Southern Hemisphere. His father was Rector of St. Fagans near Cardiff, where he was born in 1858. His mother was a Canadian and related to the author, Maria Edgeworth; hence Professor David's unusual name. His father had made a notable collection of fossils and his cousin, W. Ussher of the Geological Survey, encouraged him to study geology. At Oxford he obtained a First in Classics, but he also attended Prestwich's and later Judd's lectures on geology. His first research dealt with the glaciation of South Wales—an omen of his invaluable polar work later.

In 1881 the government geologist of New South Wales vanished during a field trip on the south coast. The mystery has never been cleared up, but David was appointed Assistant Geological Surveyor as the result of this loss. His first work dealt with the tin fields of New England, and his next research resulted in the discovery and reservation by the government of the Greta coal seam near Newcastle—now the most extensive coal field in the Southern Hemisphere. In 1891 he was appointed Professor of Geology at the University of Sydney.

In 1896 the Royal Society sent out an expedition to bore through a coral atoll to test the validity of Darwin's conclusions. The first attempt failed, and David led a second party from Sydney, largely financed by himself, which resulted in the bore reaching a depth of 1,114 feet. Much of our knowledge of coral reefs is based on this research. He was elected to the Royal Society of London in 1900.

The writer entered David's classes as a freshman after the Funafuti Expedition and is proud to think that he was in close touch with him, first as a junior teacher and later as a colleague, during the rest

[1] The *Journal of Geology* deeply regrets the passing of a highly esteemed associate editor and wishes to express its appreciation and indebtedness in this memorial.—THE EDITOR.

of his strenuous life. Each expedition he made colored David's teaching until replaced by a new effort. His lectures gave a vivid picture of "living" science, for they were based on his own major contributions as much as on current texts. Journeys in the Deccan helped to strengthen his glacial correlations, and his Antarctic experiences resulted from a desire to see an "Ice Age in Being."

Although he originally intended to return with the "Nimrod," he stayed in Antarctica during the whole of Shackleton's first expedition, 1907–8–9. On this expedition he led the first ascent of the volcano of Erebus (13,000 feet) and on January 6, 1909, he discovered the South Magnetic Pole. Only a fellow "Antarcticker" can adequately appreciate the dangers involved in that coastal journey for 200 miles on breaking pack ice, which enabled David and his party to reach the Antarctic Plateau and ultimately accomplish this notable feat. Every scientist with the slightest interest in glacial geology knows the huge memoir by David and Priestley—splendidly illustrated with hundreds of photographs and admirable sketches and diagrams—dealing with the glaciology, physiography, stratigraphy, and tectonic geology of South Victoria Land.

At the outbreak of the World War David, although nearly 60, devoted himself to raising a mining battalion which he accompanied as major. He was soon appointed Chief Geologist to the British army, and superintended problems of tunnelling, water supply, and the like throughout the war. A fall down a shaft resulted in a wound from which he suffered the rest of his life. He received the D.S.O., and in 1920 was awarded a knighthood.

He retired from the University of Sydney as Professor Emeritus in 1924, but still maintained his old rooms and carried out research as strenuously as ever. As a lecturer David has rarely been surpassed. He had a magnetic personality and a clarion voice. He gave great pains to a lecture, often spending an hour beforehand covering huge blackboards with elaborate maps and sections in colored chalks. Before him was a long and broad table covered with specimens, and often some working model to illustrate physical phenomena would be operating under the care of an attendant. Every available corner of his rooms was filled with books, specimens, and sketches, so that it was difficult to find seating accommodation.

Though he did not appear a strong man, his slight, wiry frame would cover the country at a jog-trot far ahead of most of his young students. With his huge leather bag and old mackintosh, he became a familiar sight over much of Australia. His energy and enthusiasm started many a young man on his life's work in science, and his courtesy to high and low was equally remarkable. Since his retirement he worked steadily on his four-volume geology of Australia—of which the large map has recently been published. His discovery of relatively well-preserved fossils in the pre-Cambrian of South Australia—including fairly large Eurypterids—delayed publication of this large work, but the reviewer understands that it was finished a few weeks before his death. So well did the Australian people appreciate the work of their chief scientist that Sir Edgeworth was accorded a state funeral with full military honors.

REVIEWS

Geological Researches in the Judean Desert. By L. PICARD, with an Introduction by J. W. GREGORY. Jerusalem, 1931. Pp. 108+ viii; pls. 6; map 1.

The investigations which form the basis of this monograph were conducted at the instance of the Palestine Mining Syndicate to determine the extent and quality of the phosphate and bituminous limestone deposits of the area. The stratigraphy from the Cenomanian-Turonian, the oldest-known beds, to the Pleistocene is detailed. The Cretaceous was a period of uplift or transition from marine to shallow-water lagoonal deposits, but the Cenozoic beds were all deposited under non-marine conditions.

To the structural geologist the area is of intense interest. The Jordan–Dead Sea Valley, the Red Sea, and the African Rift valleys are so often linked together in origin that clues to the genesis of one may be helpful in solving the problems connected with the others.

Laret (1869) and Hull (1886) thought that the east wall of the Jordan–Dead Sea trough was due to faulting, but that the west wall was a downfold. Faas (1867) called the structure a "graben," and has been followed by Suess (1891), Blackenthorn (1914), and Gregory (1921). Willis (1928) called it a "ramp valley."

Picard found three main systems of faults: (1) north–south, the oldest, with a "throw" of over 2,000 feet; (2) northwest–southeast, crescentic cross faults with a "throw" of less than 500 feet; and (3) north–northeast, south–southwest "Somalic dislocations," having a "throw" of 100–200 feet. The last group, named after faults with a parallel trend in British Somaliland, indicates the author's belief in genetic relationship between East African and Jordan–Dead Sea structures. It is interesting to note that the other two systems are essentially parallel to the "Red Sea" and "East African" trends of MacFadyen, whose report on the Somaliland protectorate postdated Picard's publication by two years. Picard found no evidence of ramp faults in the area. He disposed of Laret's hypothesis by discovering undisputable evidence of faulting along the western side of the valley. His theory of genesis is quite similar to that proposed by Gregory for the Great Rift valleys of East Africa.

It is indeed unfortunate that the monograph is not more clearly written, for it contains a great deal of information valuable both to the stratigra-

pher and to the structural geologist. The author sees fit to criticize Willis for his use of the term "overthrust," though he overlooks completely his own obscurity of terminology and sometimes contradictory usage.

H. W. STRALEY, III

The Geology of the Country around Reigate and Dorking. By H. G. DINES and F. H. EDMUNDS. With notes by H. DEWEY and C. J. STUBBLEFIELD and a chapter on "Paleontology" by C. P. CHATWIN. London: "Memoir Geological Survey England and Wales," 1933. Pp. vii+194; figs. 15; pls. 5. 4*s*.

This volume is explanatory of Geological Map Sheet 286 already published. The hard-rock formations from the Wealden (Lower Cretaceous) to the Bracklesham beds (Eocene) and the surficial Quarternary deposits are described in detail.

R. T. C.

Die Bodenschätze Deutschlands. Vol. I. By E. KRENKEL. Berlin: Bornträger, 1932. Pp. viii+301. RM. 13.80.

A short general survey of the various mineral resources of Germany comprises the beginning of this book and a detailed description follows of its brown coals and bituminous coals. A discussion of the origin of coal and a description of the deposits is given. Ample bibliographies add to its usefulness.

In the second volume will be considered coal mining, coal as a fuel and source of by-products, and petroleum, potash, and other useful deposits.

A. C. NOÉ

Die permo-karbonischen Eiszeiten. By WILHELM SALOMON-CALVI. Leipzig: Akademische Verlagsgesellschaft, 1933. Pp. vi+156.

An attempt is made in this book to present a critical examination of Wegener's theory in the light of all available information upon the Permo-Carboniferous glaciation. The author accepts the assumption of continental drift but refutes the idea of a continuous westward movement and flight from the poles. On the basis of the distribution of the Glossopteris floras he thinks the South Pole was covered by sea at the time of the Permo-Carboniferous glaciation. Around this polar sea were grouped South Africa, India, Australia, Antarctica, and South America. Here is another effort to adjust the position of land masses to the facts of plant geography.

A. C. NOÉ

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